



Allan Rodhe

The Origin of Streamwater Traced by Oxygen-18

Appendix

Uppsala Universitet
Naturgeografiska institutionen
Avdelningen för Hydrologi



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Appendix

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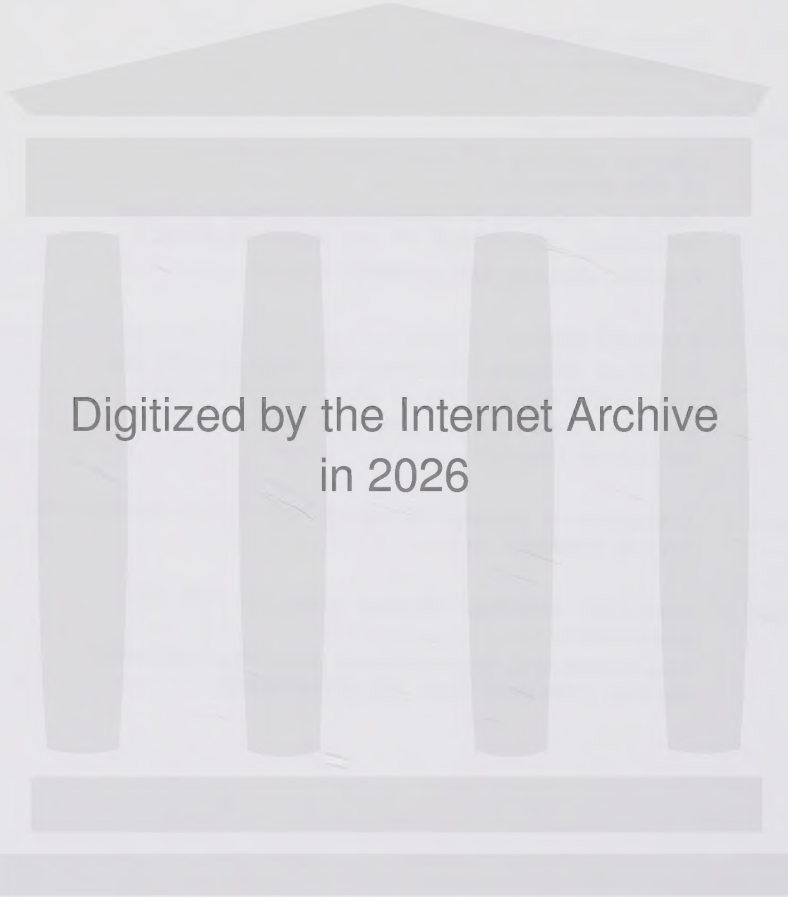
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APPENDIX 1

Figures showing $\delta^{18}\text{O}$ of precipitation and streamwater

The figures are arranged according to the name of the basin and the year of investigation. The figures show:

$\delta^{18}\text{O}$ (center of the circles) and amount (area of the circles) of daily precipitation samples, $\delta^{18}\text{O}$ of streamwater, $\delta^{18}\text{O}$ of snowpacks in late winter and spring, and stream hydrograph based on daily mean values. The symbols by which the investigated events are referred to in the text are marked in the hydrographs by Roman numerals (snowmelt events) and capital letters (rainfall events).

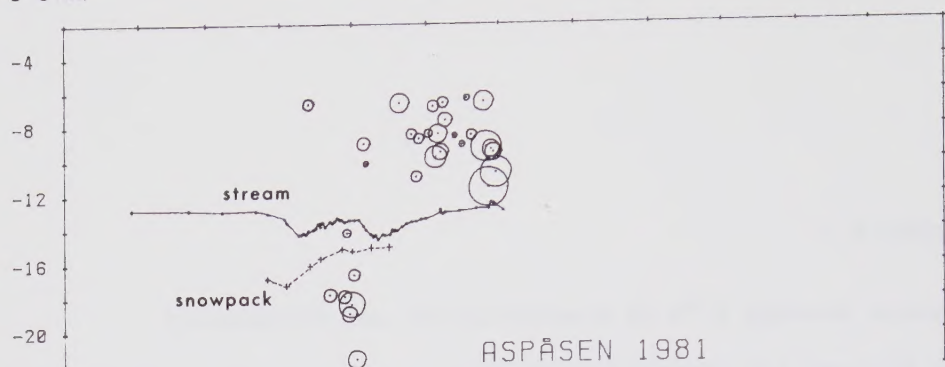
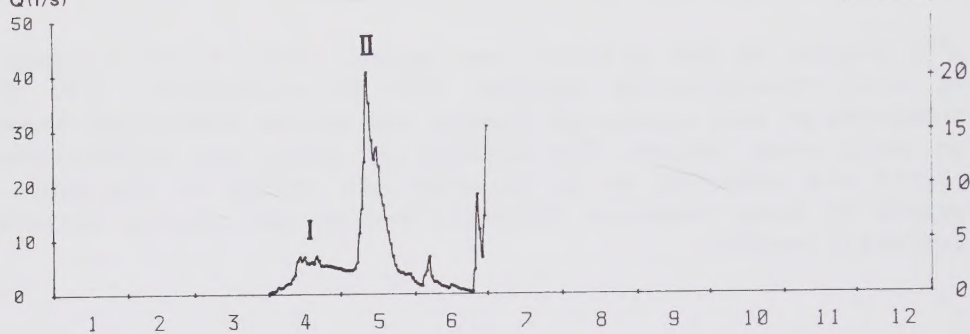
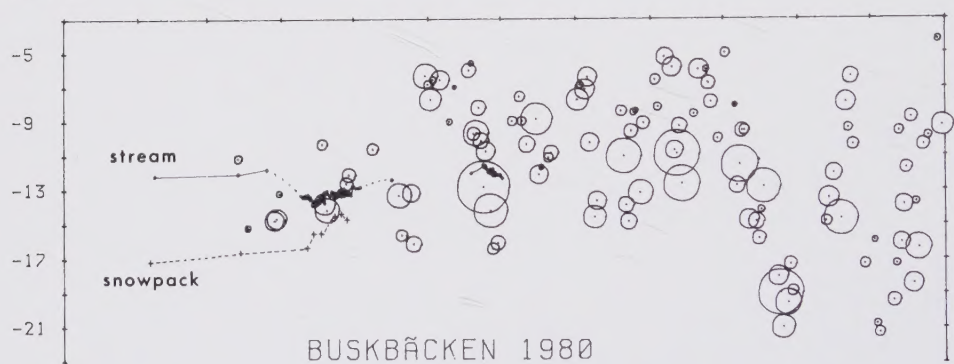
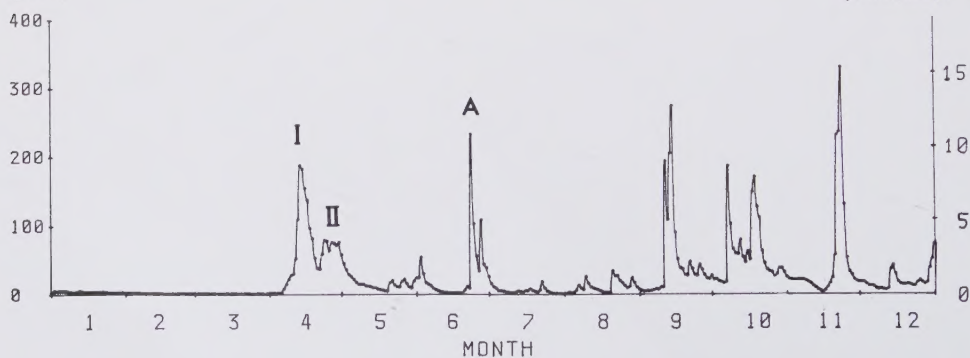
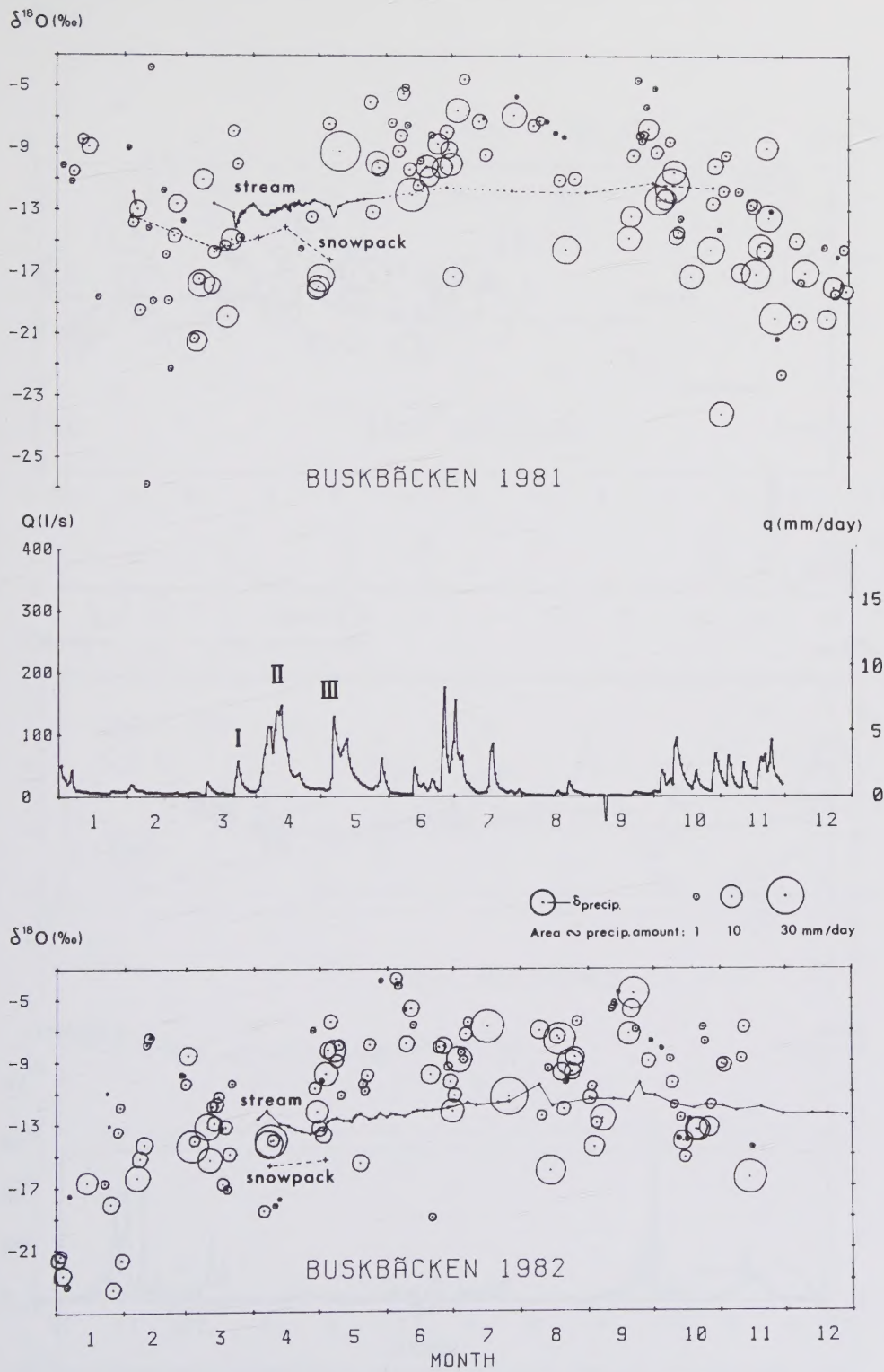
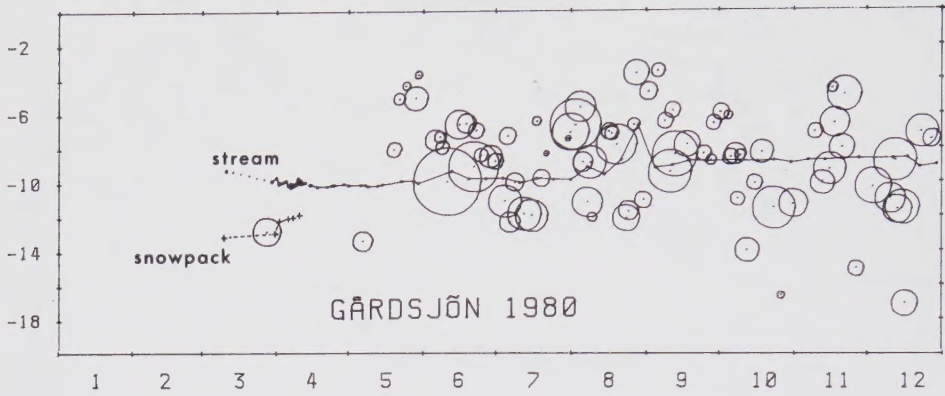
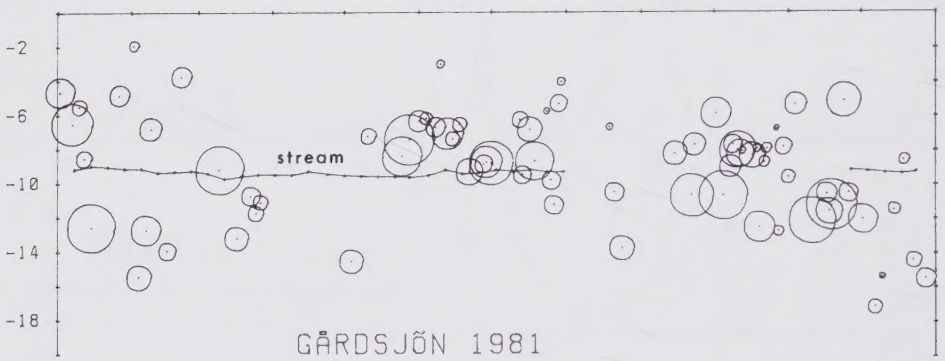
$\delta^{18}\text{O}$ (‰) Q (l/s) q (mm/day) $\delta^{18}\text{O}$ (‰) Q (l/s) q (mm/day)

Fig.A1.2 Buskbäcken 1981 and 1982.



$\delta^{18}\text{O}$ (‰) $\delta^{18}\text{O}$ (‰) Q (l/s)20
15
10
5
0 q (mm/day)40
30
20
10
0

1 2 3 4 5 6 7 8 9 10 11 12

MONTH

Fig.A1.4 Gårdsjön F1 1982 and comparisons between Gårdsjön F1 and F2 1980 - 1982 (enlarged $\delta^{18}\text{O}$ scales).

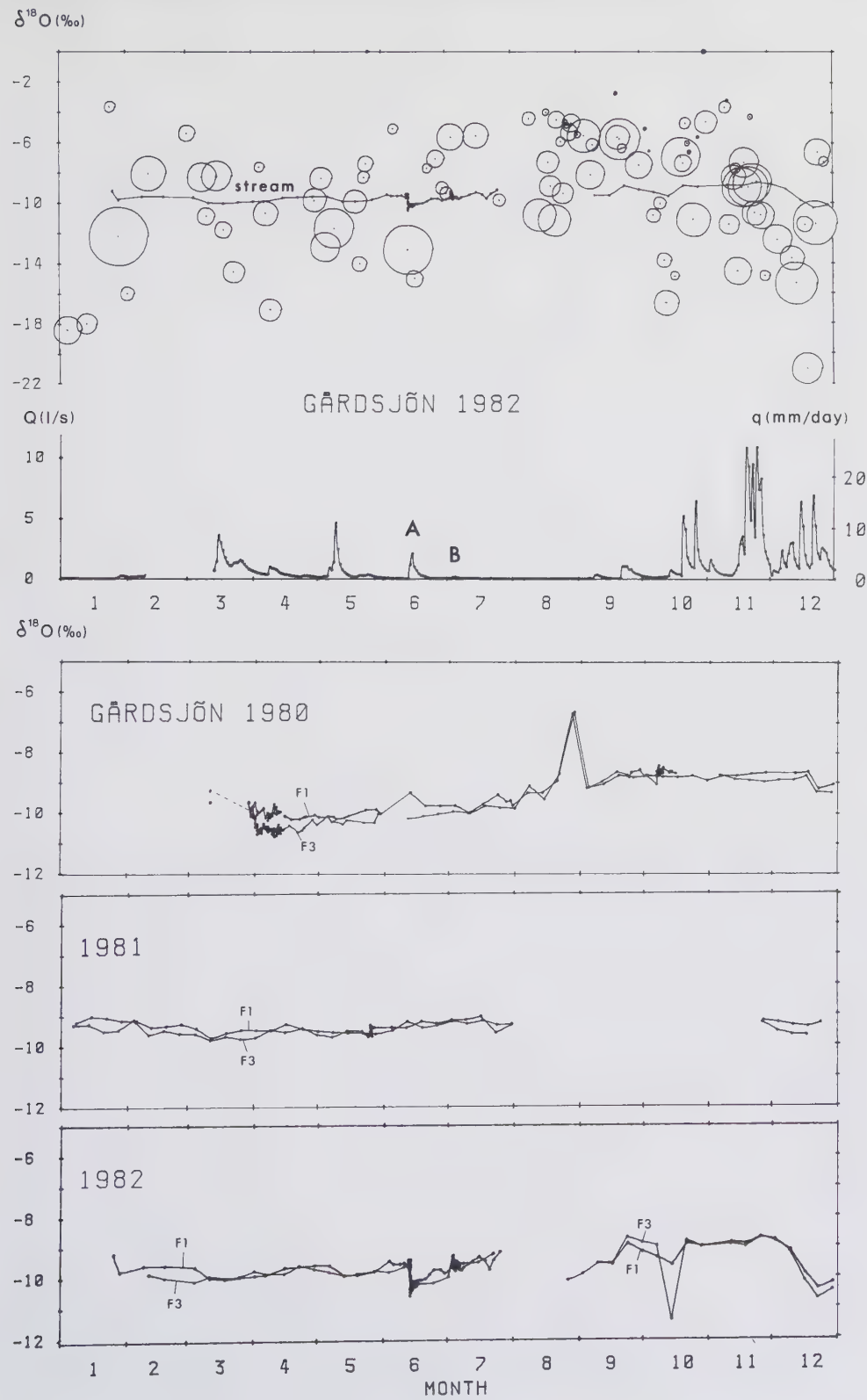
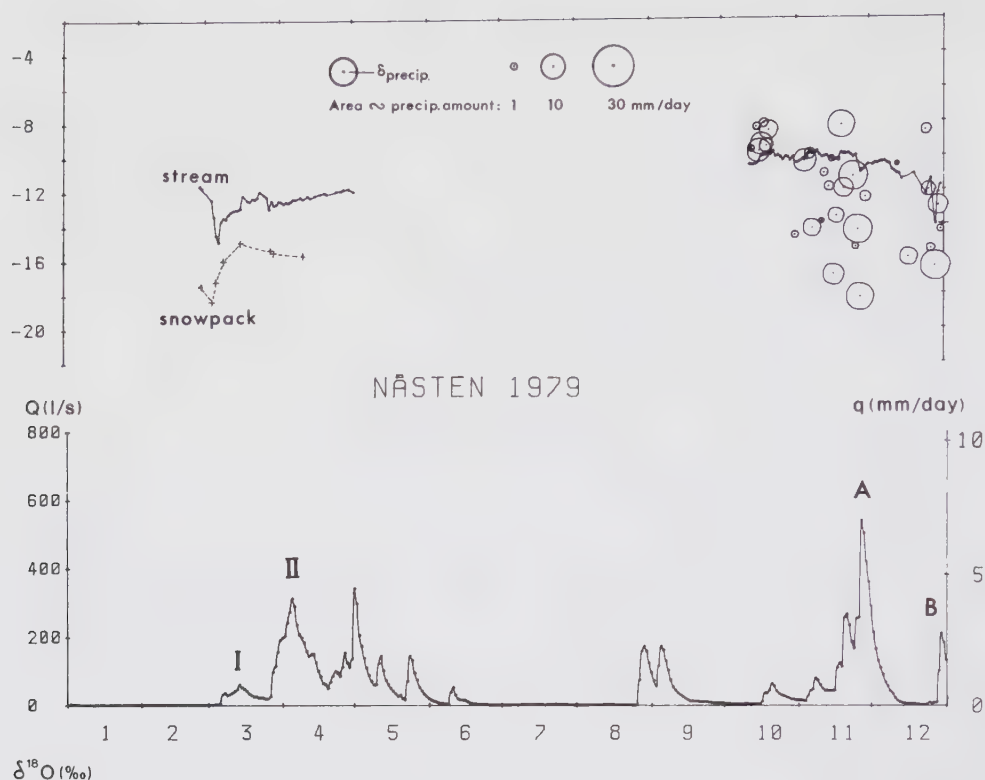


Fig.A1.5 Nästen 1979 and 1980. $\delta^{18}\text{O}$ of precipitation not observed before October 1979.

$\delta^{18}\text{O}$ (‰)



$\delta^{18}\text{O}$ (‰)

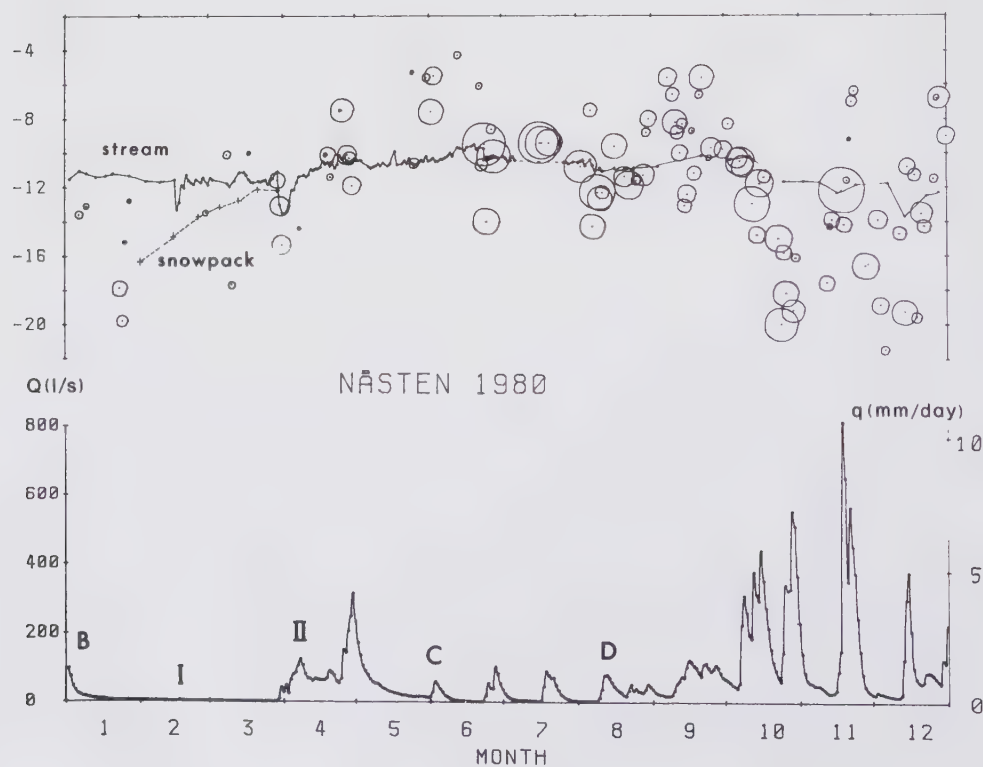
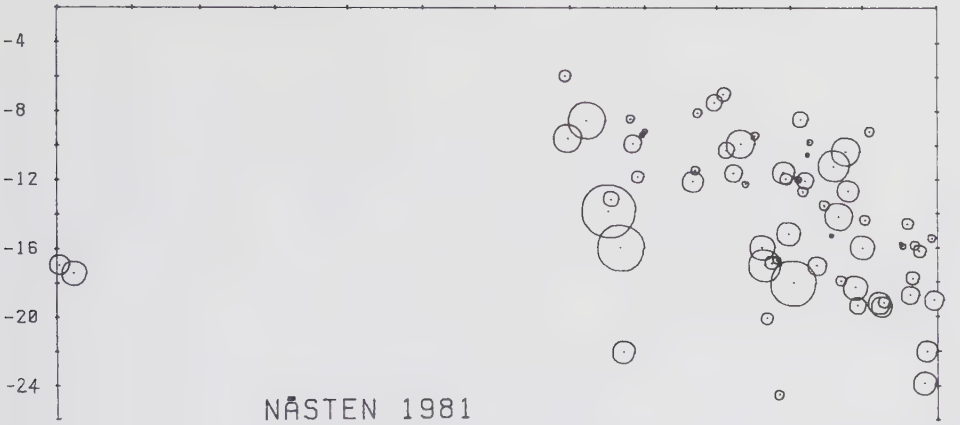


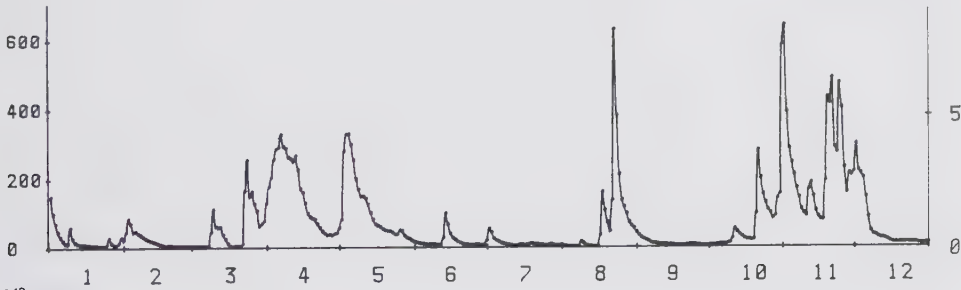
Fig.A1.6 Nāsten 1981 and 1982.

$\delta^{18}\text{O}$ (‰)

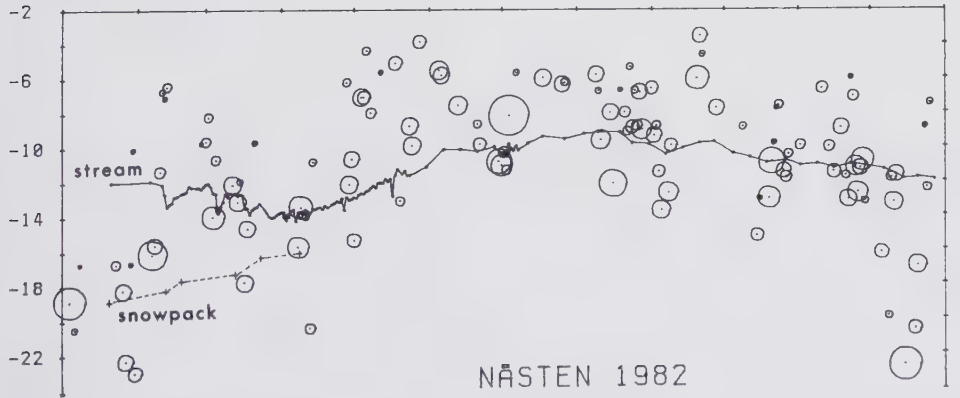


Q (l/s)

q (mm/day)



$\delta^{18}\text{O}$ (‰)



Q (l/s)

q (mm/day)

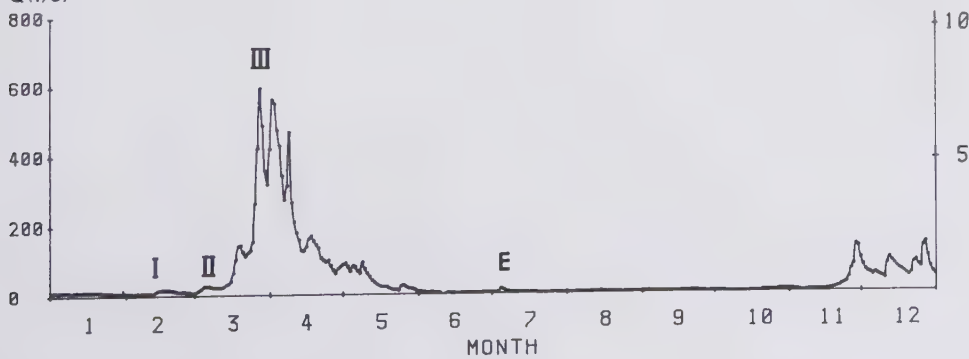
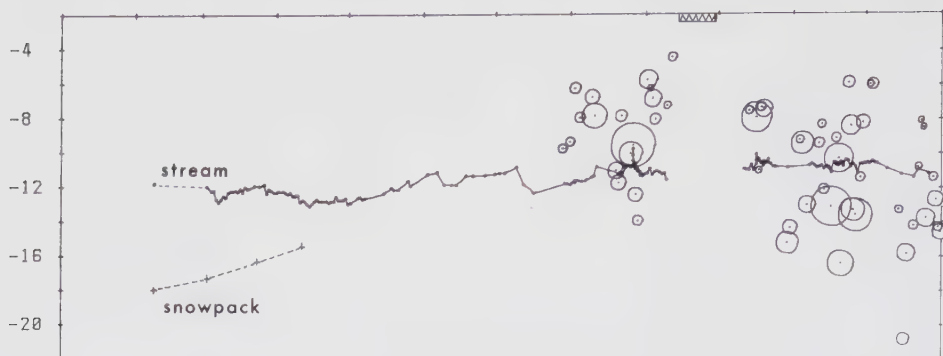


Fig.A1.7 Stormyra 1979 and 1980. $\delta^{18}\text{O}$ of precipitation not observed before August 1979 and during the second half of September 1979, zig-zag line.

$\delta^{18}\text{O}$ (‰)



STORMYRA 1979

Q (l/s)

800

600

400

200

0

q (mm/day)

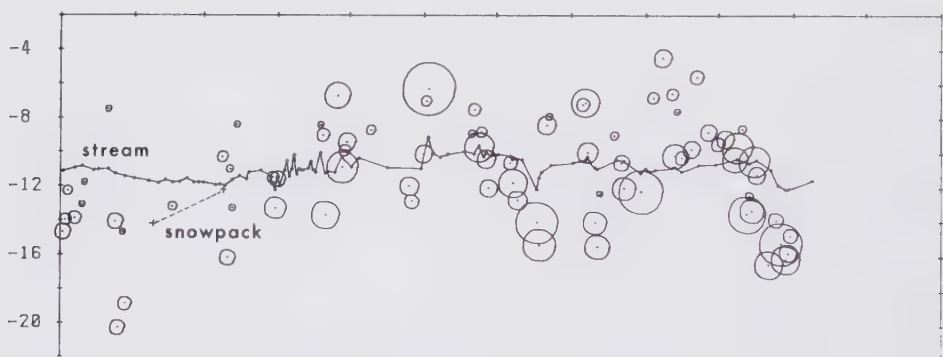
15

10

5

0

$\delta^{18}\text{O}$ (‰)



STORMYRA 1980

Q (l/s)

800

600

400

200

0

q (mm/day)

15

10

5

0

MONTH

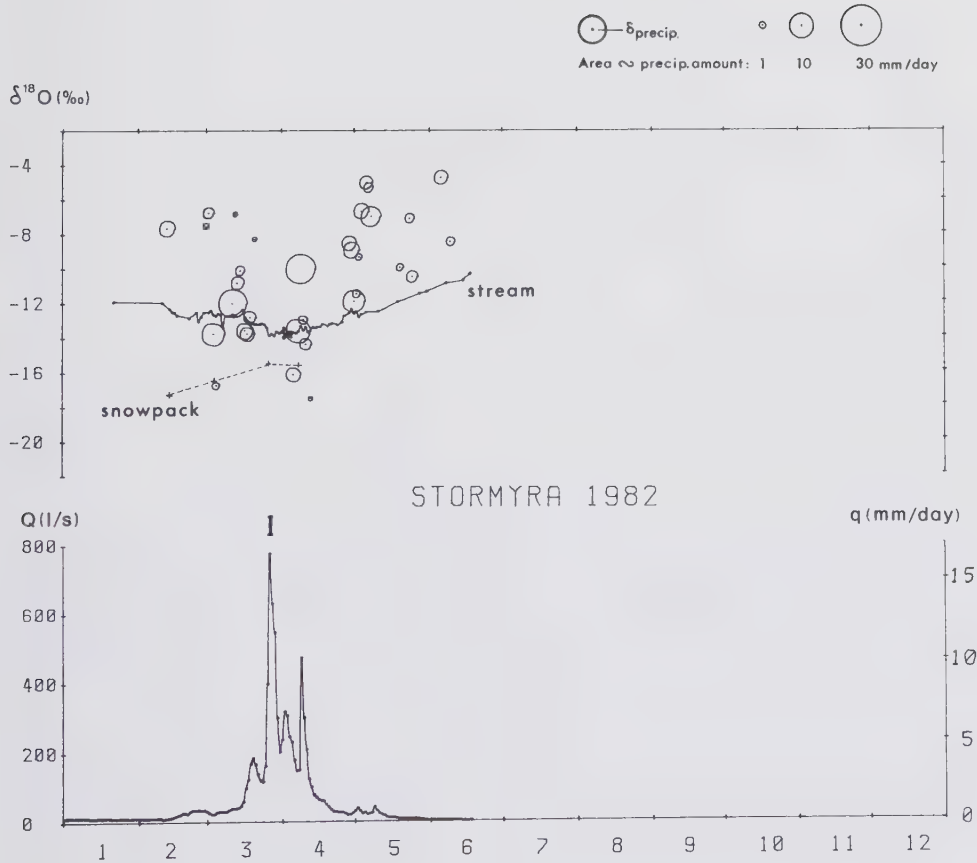


Fig.A1.9a Svartberget Nedre 1981 and comparisons between Svartberget Nedre (N) and Övre (Ö) 1981 (enlarged $\delta^{18}\text{O}$ scales). $\delta^{18}\text{O}$ of precipitation not observed from mid June to mid August 1981, zig-zag line.

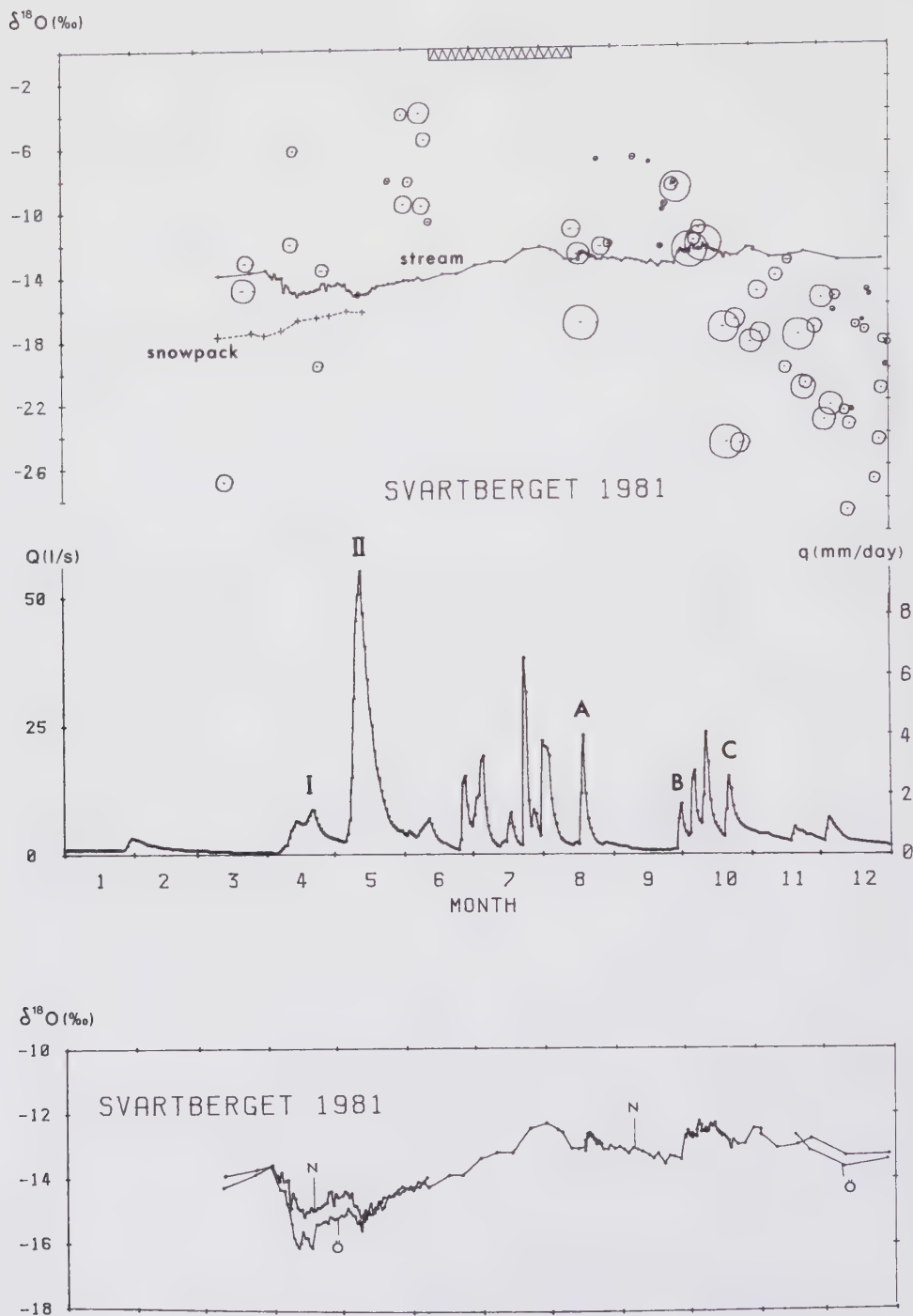
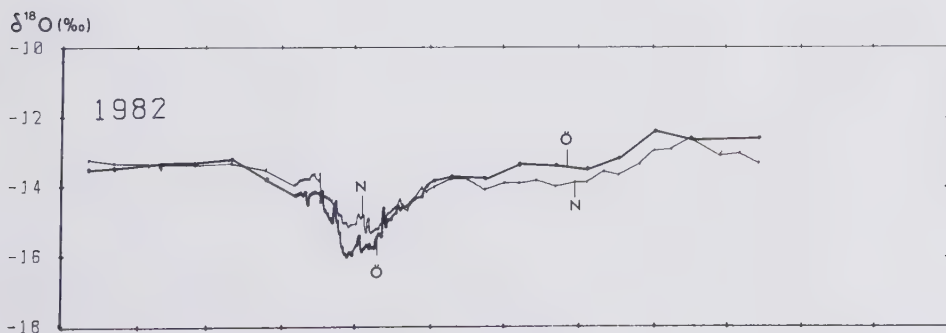
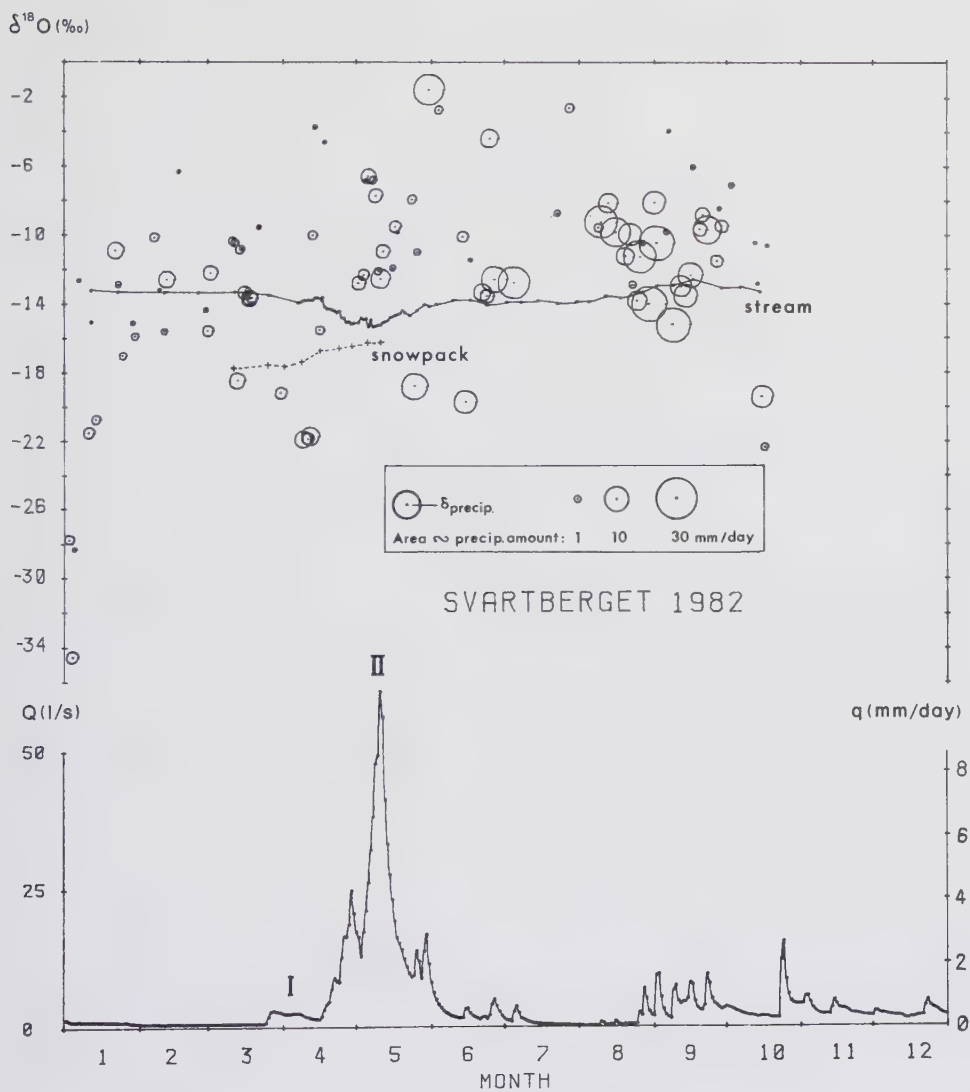


Fig.A1.9b. Svartberget Nedre 1982 and comparisons between Svartberget Nedre (N) and Övre (Ö) 1982 (enlarged $\delta^{18}\text{O}$ scales).

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APPENDIX 2 SNOWMELT EVENTS

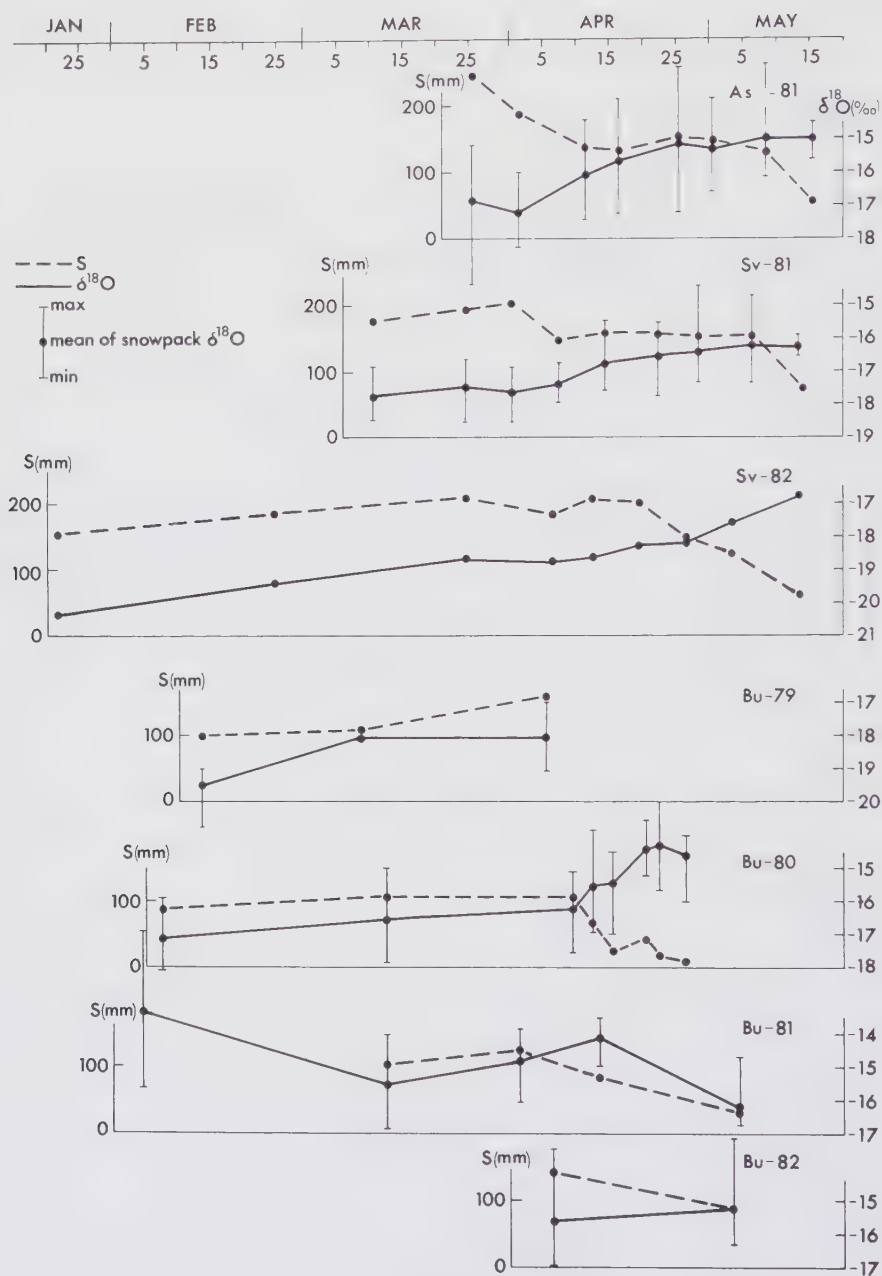
Figures showing $\delta^{18}\text{O}$ and water equivalent of the snowpacks

Table on results of hydrograph separations and hydrologic conditions of the snowmelt events

Figures showing all separated snowmelt events

The figures are arranged according to the name of the basin and the year of investigation. The figures show:

Stream hydrograph with calculated groundwater hydrograph, $\delta^{18}\text{O}$ of groundwater (δ_g) calculated using various model reservoir volumes (V_g), $\delta^{18}\text{O}$ of streamwater and snowpack, amount and $\delta^{18}\text{O}$ and type of precipitation, and calculated fractions of groundwater for various values of V_g . Calculated flow of meltwater is shown in some of the figures. The periods over which summations have been performed are marked by Roman numerals.

Fig.A2.1 Measured $\delta^{18}\text{O}$ and water equivalent of the snowpacks.

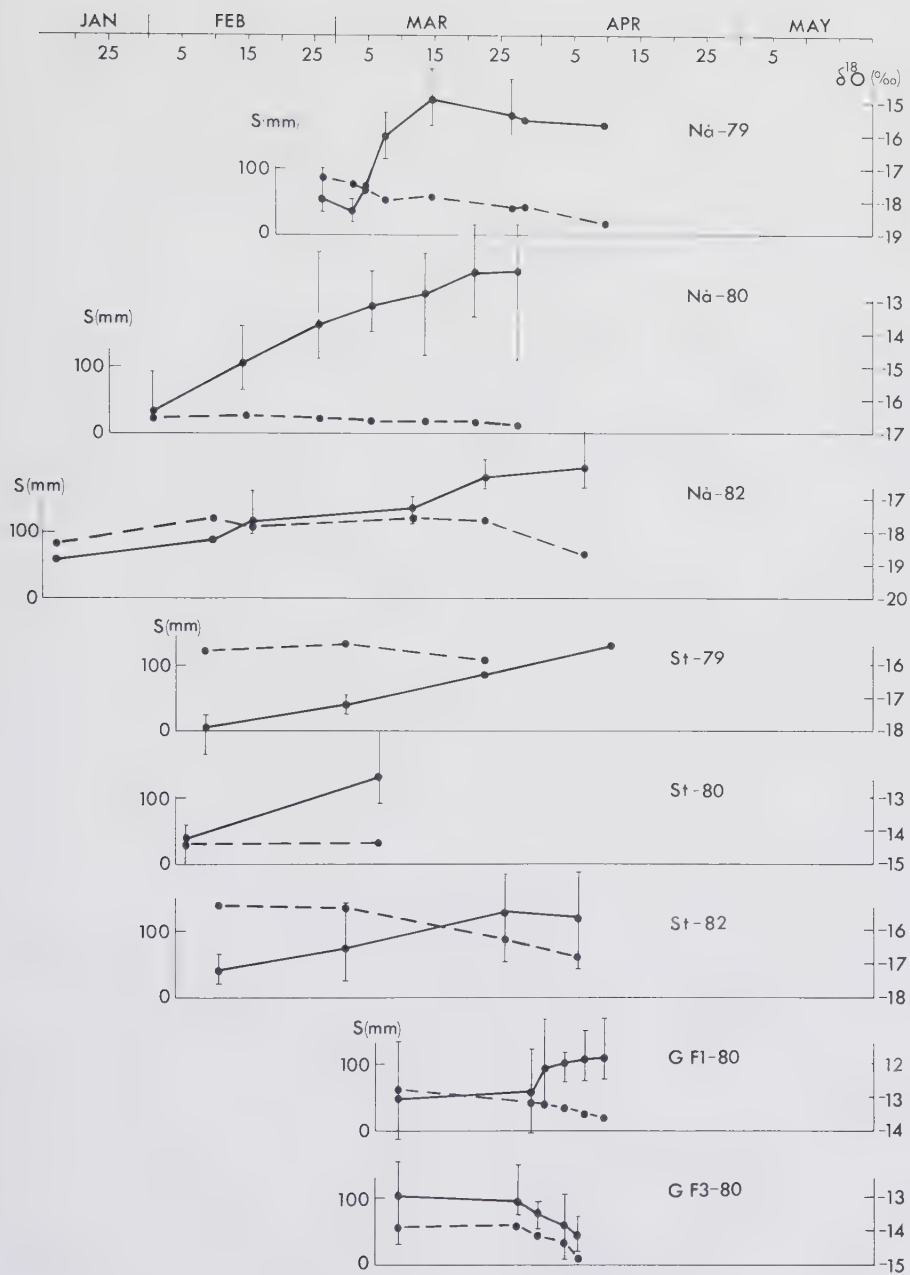
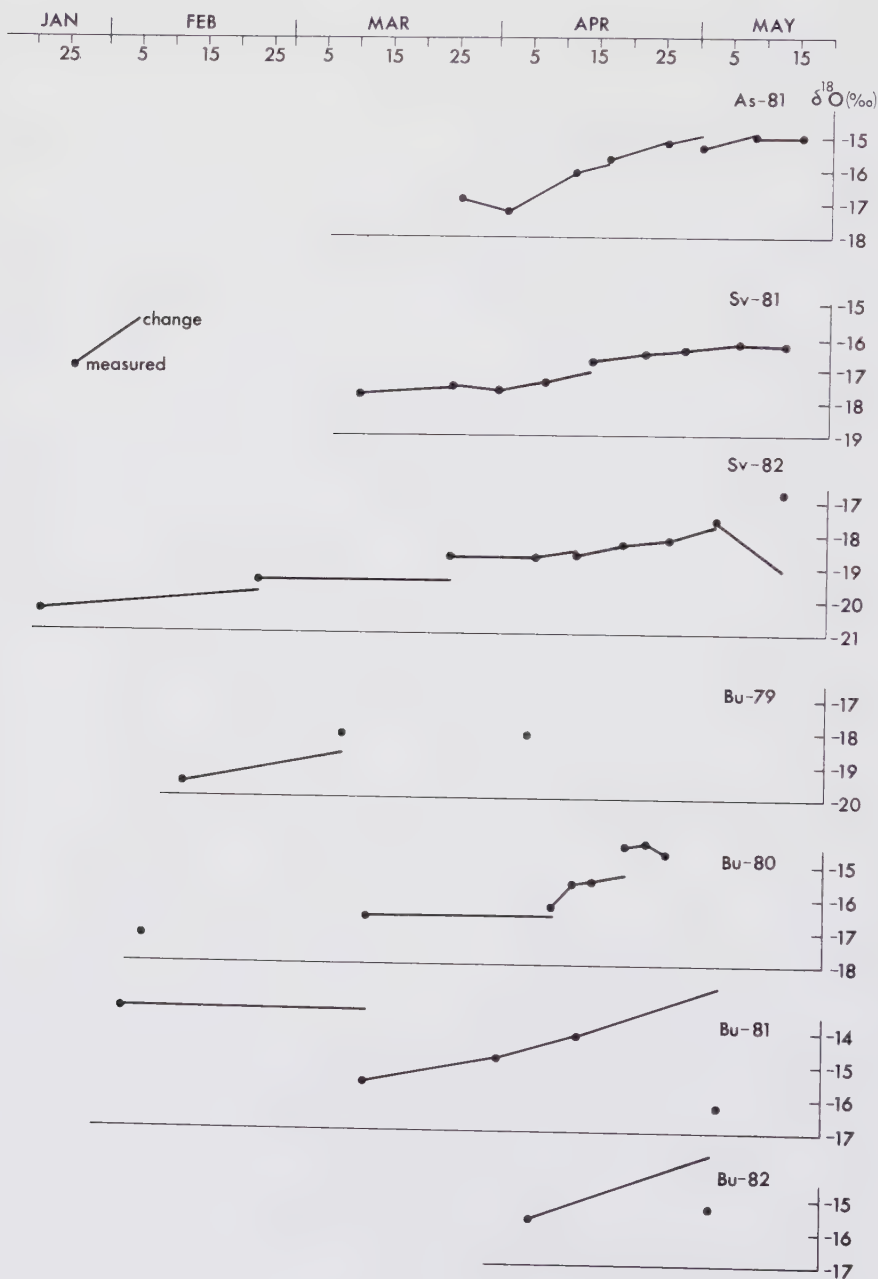


Fig.A2.2 Measured $\delta^{18}\text{O}$ of snowpacks and calculated changes of snowpack $\delta^{18}\text{O}$ after elimination of the isotopic input by precipitation.



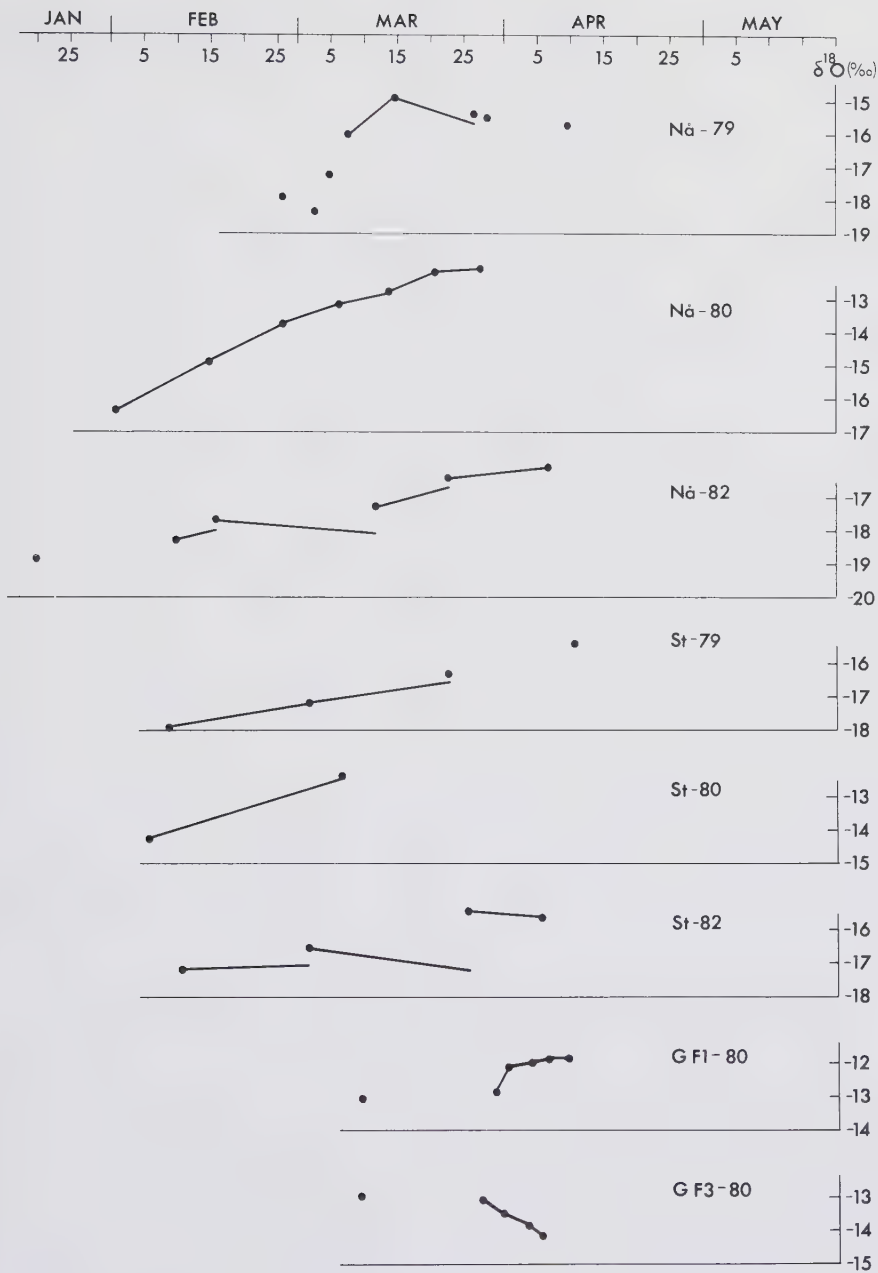


Table A2.1 Results of hydrograph separation: snowmelt events.

Basin		Period	Change in snow storage (mm)	Precipi- tation (mm)	Total runoff ¹ (mm)	Maximum discharge (1 s ⁻¹ km ⁻²)
Aspåsen	1981 I	26/3 - 5/5	-100	32.1	51 (59)	32
	II	5/5 -19/5	-146	3.4	132 (148)	218
	total	26/3 -19/5	-246	35.5	183 (207)	
Buskbäcken	1980 I	9/4 -19/4	-60	9.4	49 (50)	128
	II	19/4 -28/4	-38	13.3	32 (30)	66
	total	9/4 -28/4	-98	22.7	80 (80)	
	1981 I	22/3 - 1/4	0	6.6	9 (9)	32
	II	1/4 - 3/5	-90	36.0	80 (78)	123
	III	3/5 - 8/5		3.4	19 (14)	12
	total	22/3 - 8/5	≈-110	46.6	108 (100)	
Gårdsjön F1	1980 total	31/3 -12/4	-30	0	42	67
Gårdsjön F2	1980 total	3/4 -12/4	-25	0	22	43
Gårdsjön F3	1980 total	29/3 -12/4	-56	0	48	71
Nåsten	1979 I	26/2 -25/3	-40	7.6	8 (8)	9
	II	25/3 -19/4	-48	17.4	54 (53)	47
	total	26/3 -19/4	-88	25.0	62 (61)	
	1980 I	9/2 - 1/3	≈0		0.1	0.2
	II	1/3 -25/4	≈-20		25	19
	total	9/2 -25/4	≈-20		25	
	1982 I	5/2 -28/2	0	22.9	3 (3)	3
	II	28/2 -13/3	0	18.7	4 (3)	5
	III	13/3 -12/4	-75	34.1	120 (108)	108
	total	5/2 -12/4	-75	75.7	127 (114)	

¹ The first figure is the runoff calculated from hydrographs based on water levels at the hours of water sampling.

² The groundwater fractions are calculated by using various values of the model reservoir V_g in Eq.(3.6). $V_g = \infty$ corresponds to the use of constant $\delta^{18}O$ of groundwater.

³ ΔX_{tot} = sensitivity in the calculated groundwater fraction X to systematical errors in the $\delta^{18}O$ of meltwater of $\pm 1\text{‰}$ and in the $\delta^{18}O$ of groundwater of $\pm 0.5\text{‰}$. See the uncertainty discussion in the main text, p. 215.

⁴ The period mean groundwater fractions are calculated assuming momentary discharge values proportional to those observed at Svartberget Nedre.

Fraction of groundwater ² (%)			Fraction of discharge area (%)			ΔX_{tot}^3	ΔY_{tot}^3
$V_g = \infty$	$V_g = 500 \text{ mm}$	$V_g = 250 \text{ mm}$	by ^{18}O $V_g = \infty$	by field $V_g = 250 \text{ mm}$	by field survey	(%)	(%)
70	73	77	12	(9)			
32	37	44	60	(49)	≈ 10		
42	47	53	37	(30)		31	23
54	57	61	32	(29)			
50	57	67	33	(22)	28		
53	57	63	35	(26)		20	13
55	55	55	63	(62)			
57	61	66	27	(22)			
68	74	80					
59	63	67	29	(23)		24	14
72	76	79	38	(30)		17	26
67	68	70	29	(27)		23	20
80	83	86	20	(13)	15	10	9
61	61	61	7	(7)			
79	83	88	18	(16)	23		
77	80	85	13	(11)		12	6
89	89	89	2	(2)			
83	83	84	3	(4)			
55	60	67	49	(36)	31		
57	62	69	36	(27)		11	11

Table A2.1 (Continued)

Basin	Period		Change in snow storage (mm)	Precipitation (mm)	Total runoff ¹ (mm)	Maximum discharge (l s ⁻¹ km ⁻²)
Stormyra	1979	I	2/3 -24/3	-25	16.4	25 (28)
		II	24/3 -12/4	-85	45.6	82 (77)
		total	2/3 -12/4	-110	62.0	108 (103)
	1980	total	13/2 -15/4	≈-20	31 (24)	43
	1982	total	10/2 - 6/4	-85	49.6	148 (145)
Svartberget Övre	1981 ⁴	I	1/4 - 4/5	-50	11.1	
		II	4/5 -20/5	-50	0	
		total	1/4 -20/5	-200	11.1	
	1982	I	25/3 -16/4	0	17.7	10
		II	16/4 -20/5	-205	26.6	134
		total	25/3 -20/5	-205	44.3	143
Svartberget Västra	1981 ⁴	I	1/4 - 4/5			
		II	4/5 -20/5			
		total	1/4 -20/5	-200		
	1982 ⁴	I	25/3 -18/4			
		II	18/4 -18/5			
		total	25/3 -18/5	-205		
Svartberget Nedre	1981	I	1/4 - 4/5	-50	11.1	21 (23)
		II	4/5 -20/5	-150	0	91 (76)
		total	1/4 -20/5	-200	11.1	111 (98)
	1982	I	25/3 -16/4	0	17.7	15
		II	16/4 -20/5	-205	26.6	107
		total	25/3 -20/5	-205	44.3	122

¹ The first figure is the runoff calculated from hydrographs based on water levels at the hours of water sampling.

² The groundwater fractions are calculated by using various values of the model reservoir V_g in Eq.(3.6). $V_g = \infty$ corresponds to the use of constant $\delta^{18}O$ of groundwater.

³ Δx_{tot} = sensitivity in the calculated groundwater fraction X to systematical errors in the $\delta^{18}O$ of meltwater of $\pm 1\text{‰}$ and in the $\delta^{18}O$ of groundwater of $\pm 0.5\text{‰}$. See the uncertainty discussion in the main text, p. 215.

⁴ The period mean groundwater fractions are calculated assuming momentary discharge values proportional to those observed at Svartberget Nedre.

Fraction of groundwater ² (%)			Fraction of discharge area (%)			Δx_{tot}^3	Δy_{tot}^3
$V_g = \infty$	$V_g = 500 \text{ mm}$	$V_g = 250 \text{ mm}$	by ^{18}O	by field		(%)	(%)
			$V_g = \infty$	$V_g = 250 \text{ mm}$	survey		
90	92	95	6	(3)			
85	98		10				
86			9			12	7
56	63	72	49	(31)	35	16	18
40							
42							
41							
91	91	92	5	(4)			
57	65	76	25	(14)			
59	67	77	23	(13)		28	
80							
56							
64							
95							
61							
64							
60	61	61	14	(13)			
44	47	51	33	(29)	25		
47	50	53	28	(25)		25	12
84	85	87	13	(12)			
46	51	57	25	(23)			
51	55	60	25	(23)		18	8

Fig.A2.3 Aspåsen 26/3 - 23/5 1981

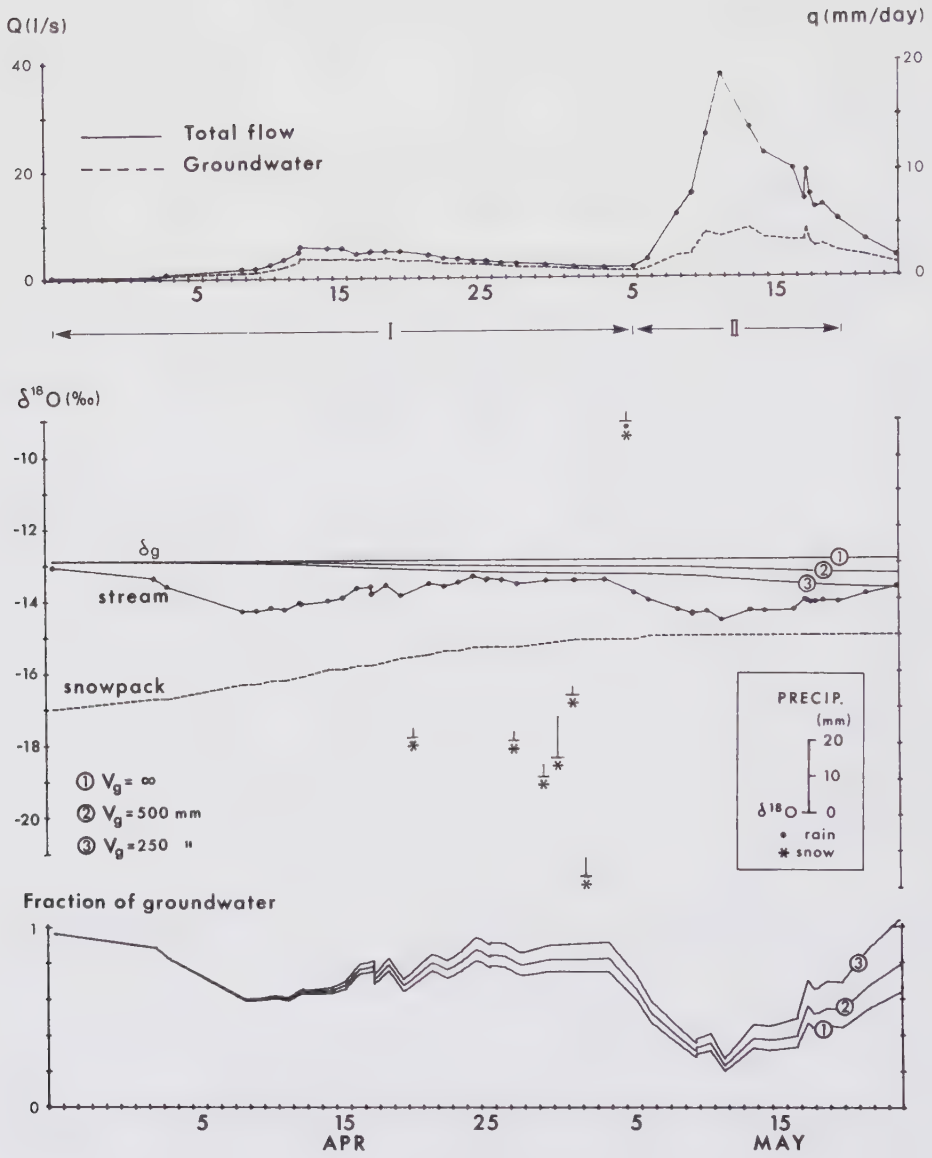


Fig.A2.4 Buskbäcken 25/3 - 28/4 1980. Legend, see Fig.A2.3.

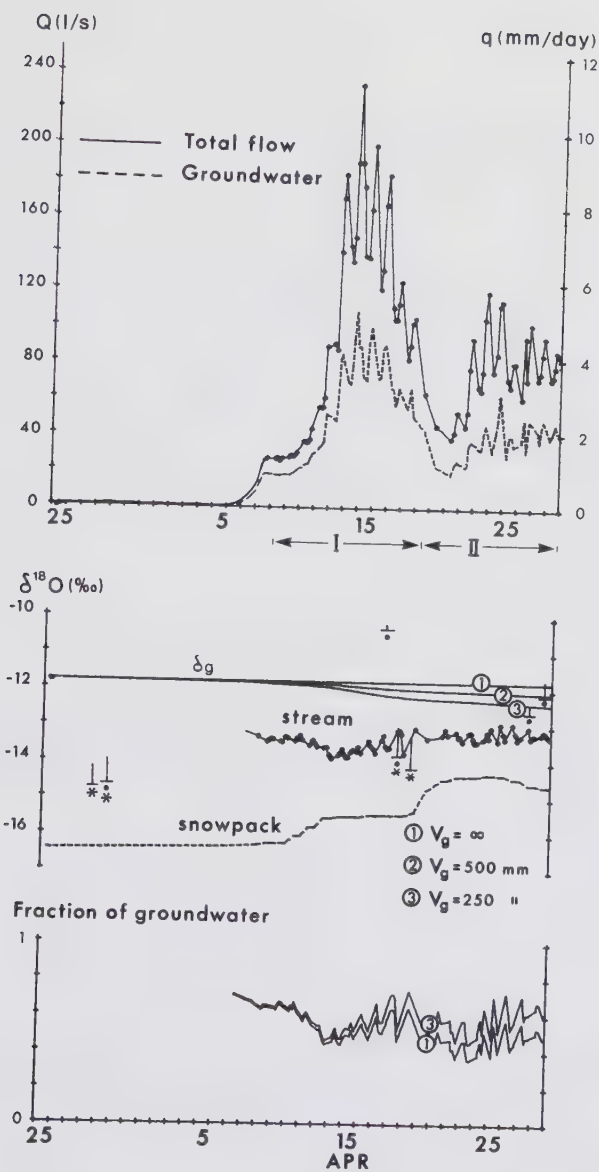


Fig.A2.5 Buskbäcken 22/3 - 8/5 1981

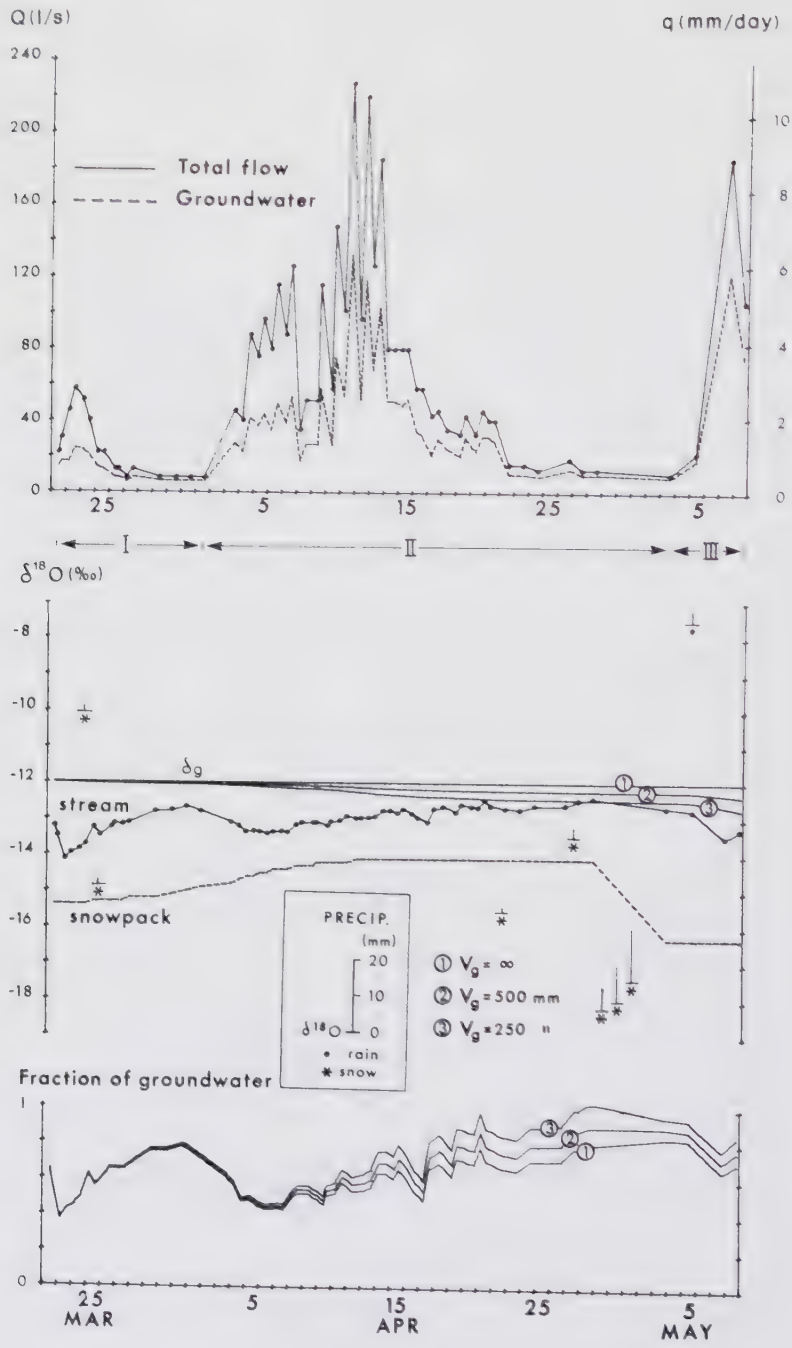


Fig.A2.6 Gårdsjön F1 31/3 -17/4 1980. Long broken line: calculated meltwater flow.

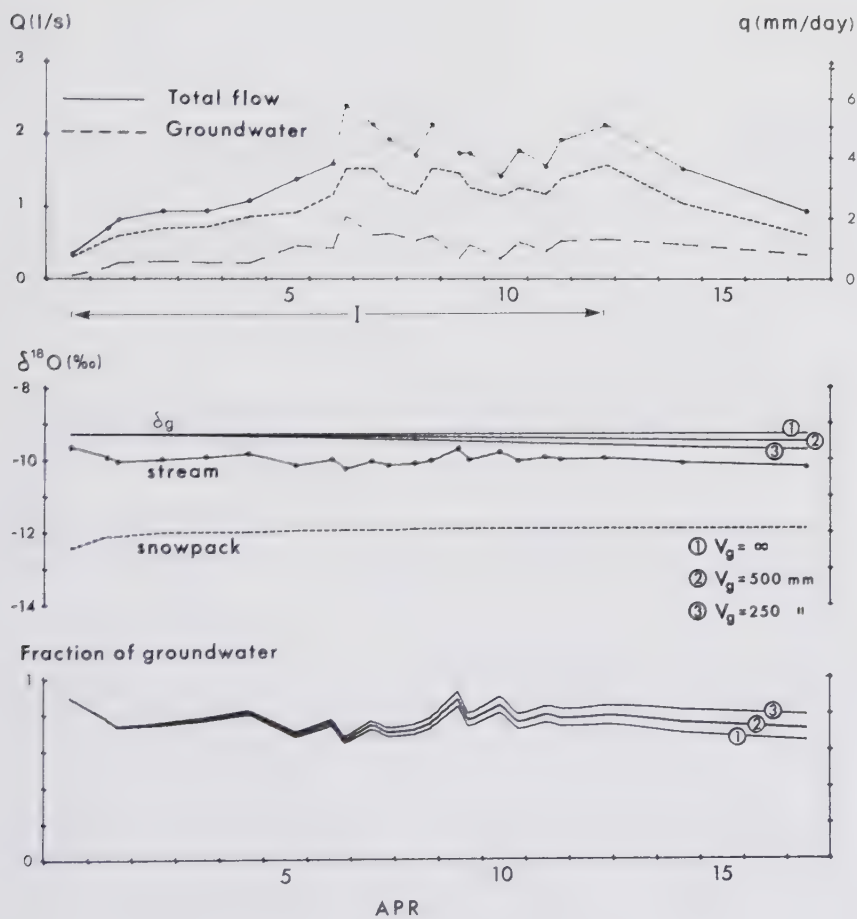


Fig.A2.7 Gårdsjön F2 3/4 - 12/4 1980. Legend, see Fig.A2.6.

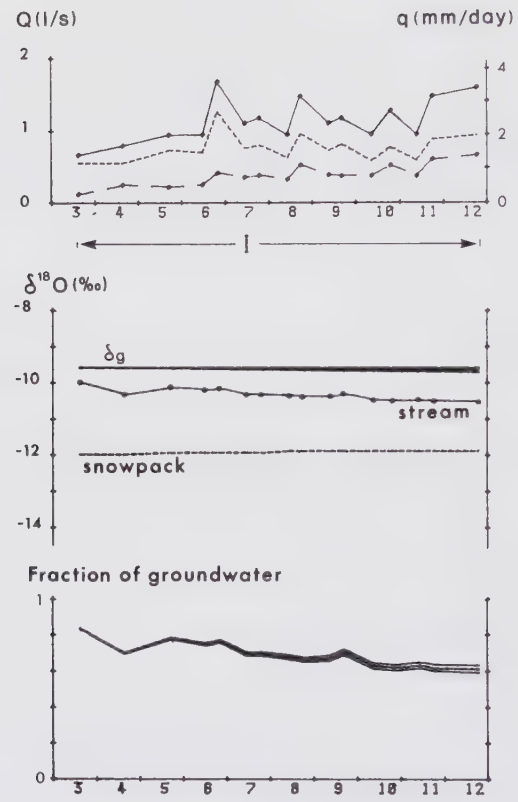


Fig.A2.8 Gårdsjön F3 29/3 -17/4 1980. Legend, see Fig.A2.6.

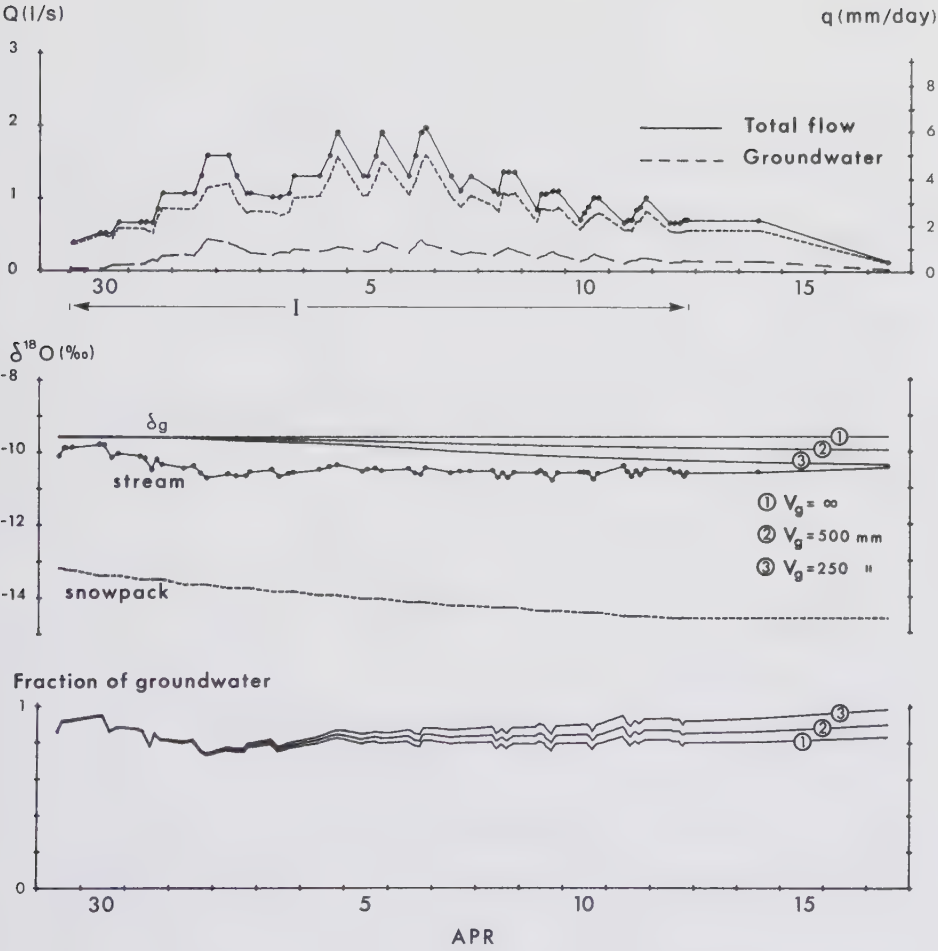


Fig.A2.9 Nåsten 26/2 - 19/4 1979.

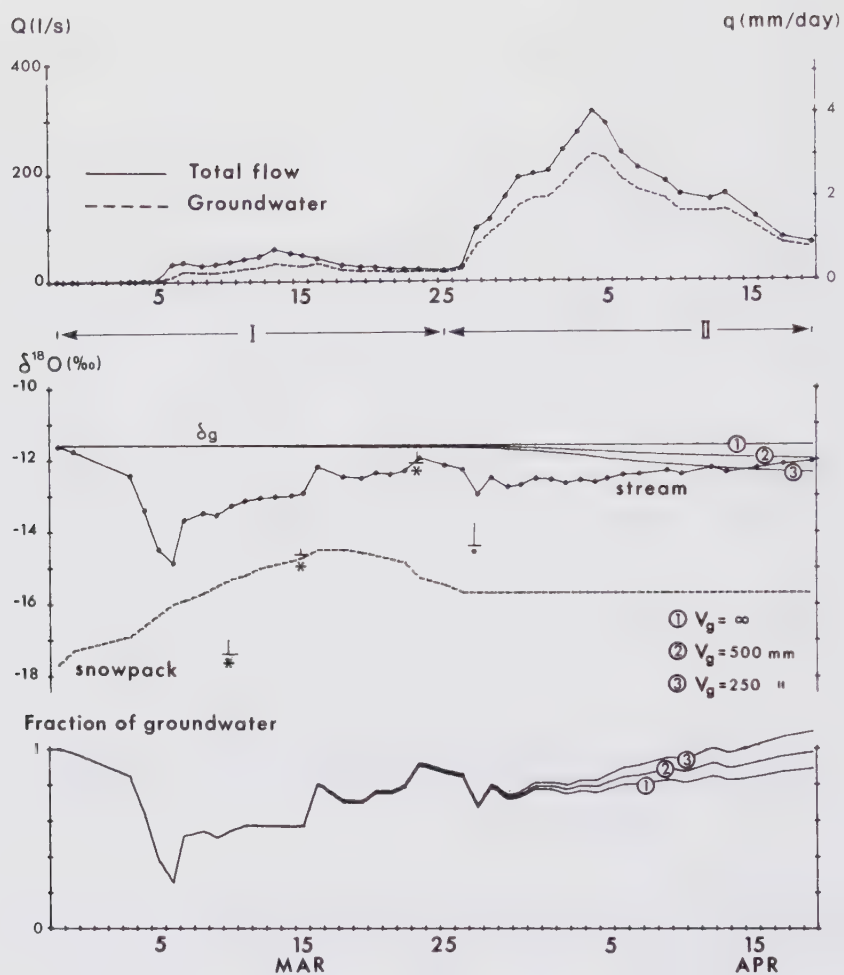


Fig.A2.10 Nâsten 9/2 - 1/3 1980 (upper diagram) and 1/3 - 25/4 1980 (lower diagram). Note the very different discharge scales.

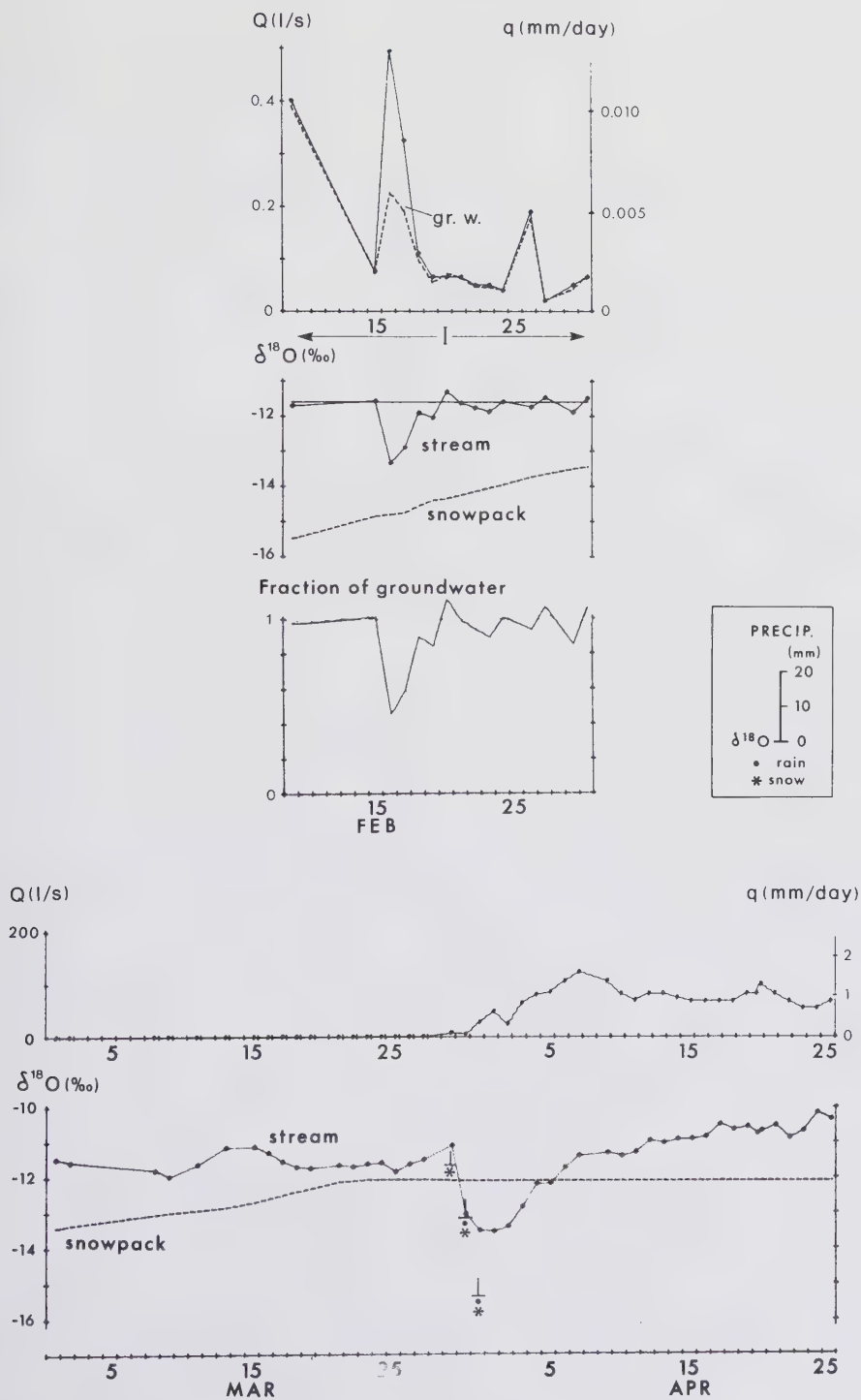


Fig.A2.11 Nåsten 5/2 - 28/2 1982.

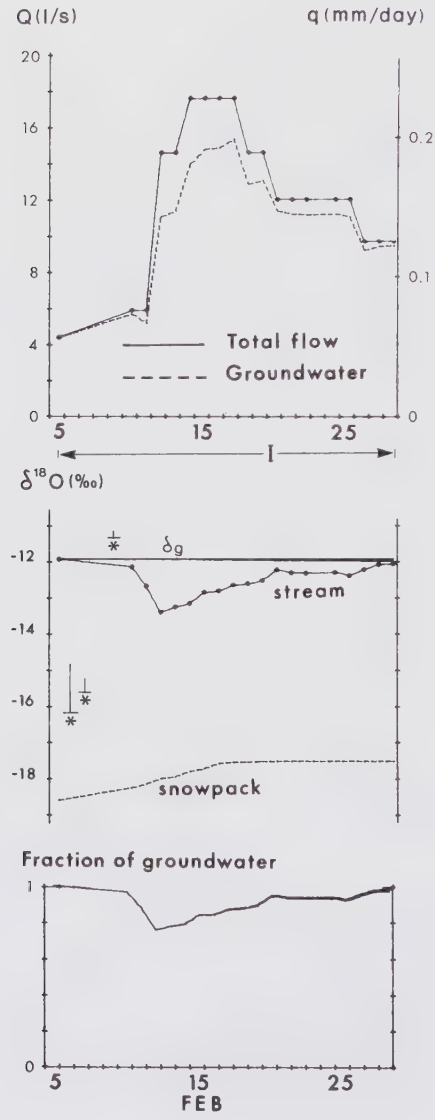


Fig.A2.12 Nåsten 28/2 - 30/4 1982.

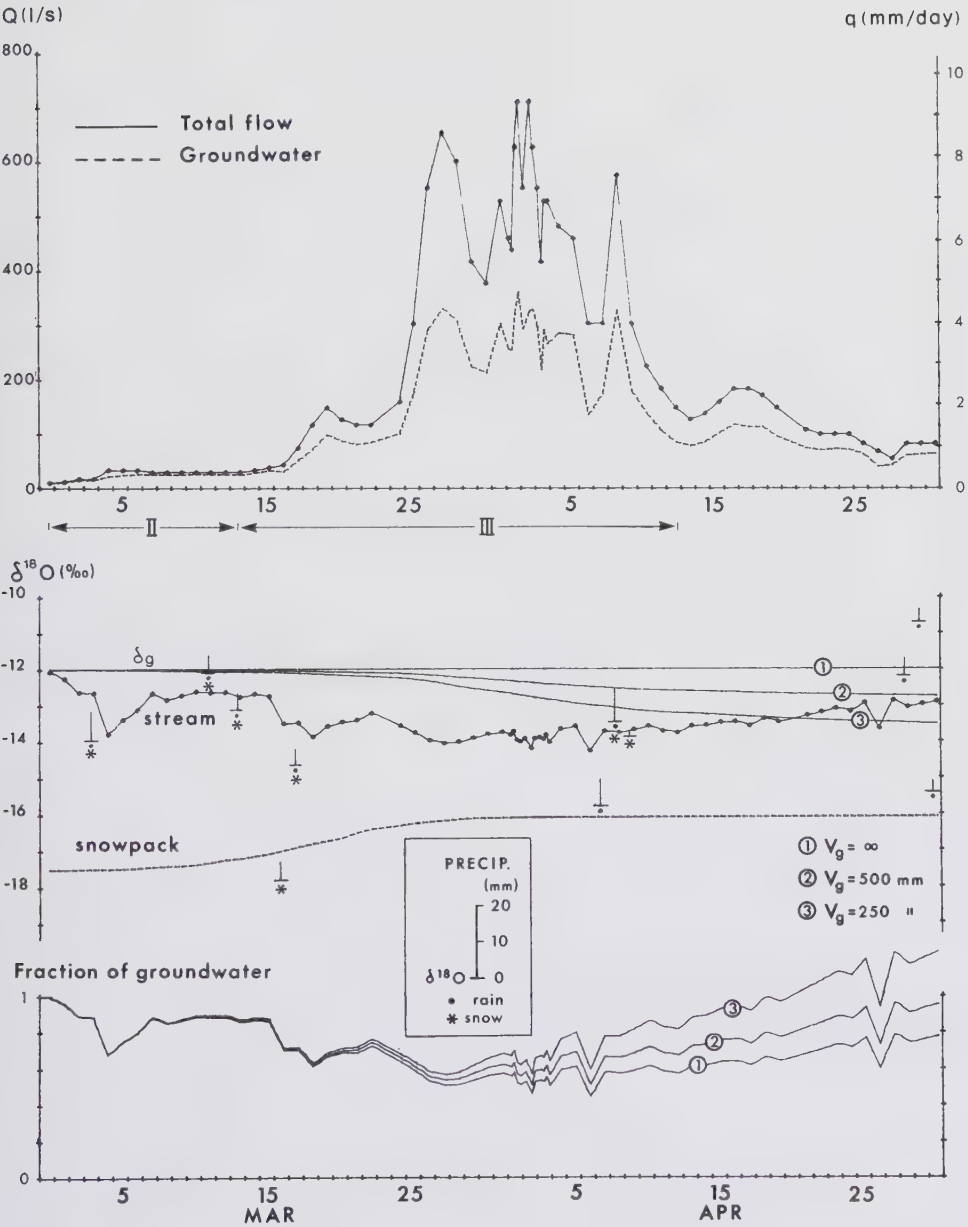


Fig.A2.13 Stormyra 2/3 - 18/4 1979. $\delta^{18}\text{O}$ of precipitation not observed.

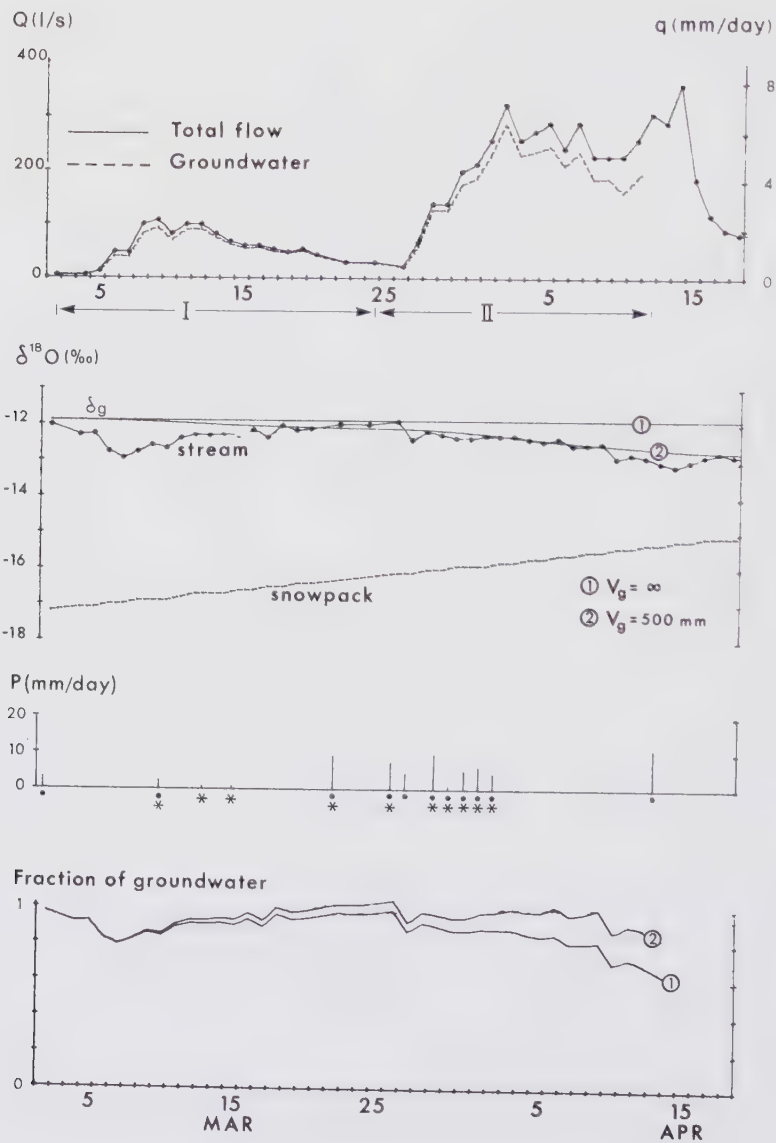


Fig.A2.14 Stormyra 13/2 - 15/4 1980.

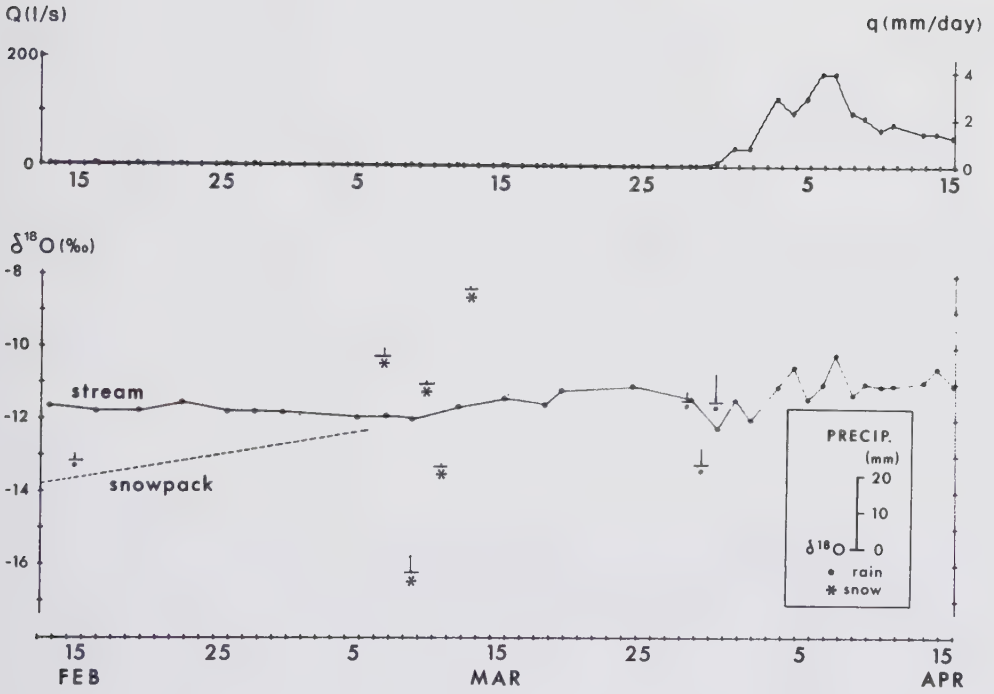


Fig.A2.15 Stormyra 10/2 - 15/4 1982.

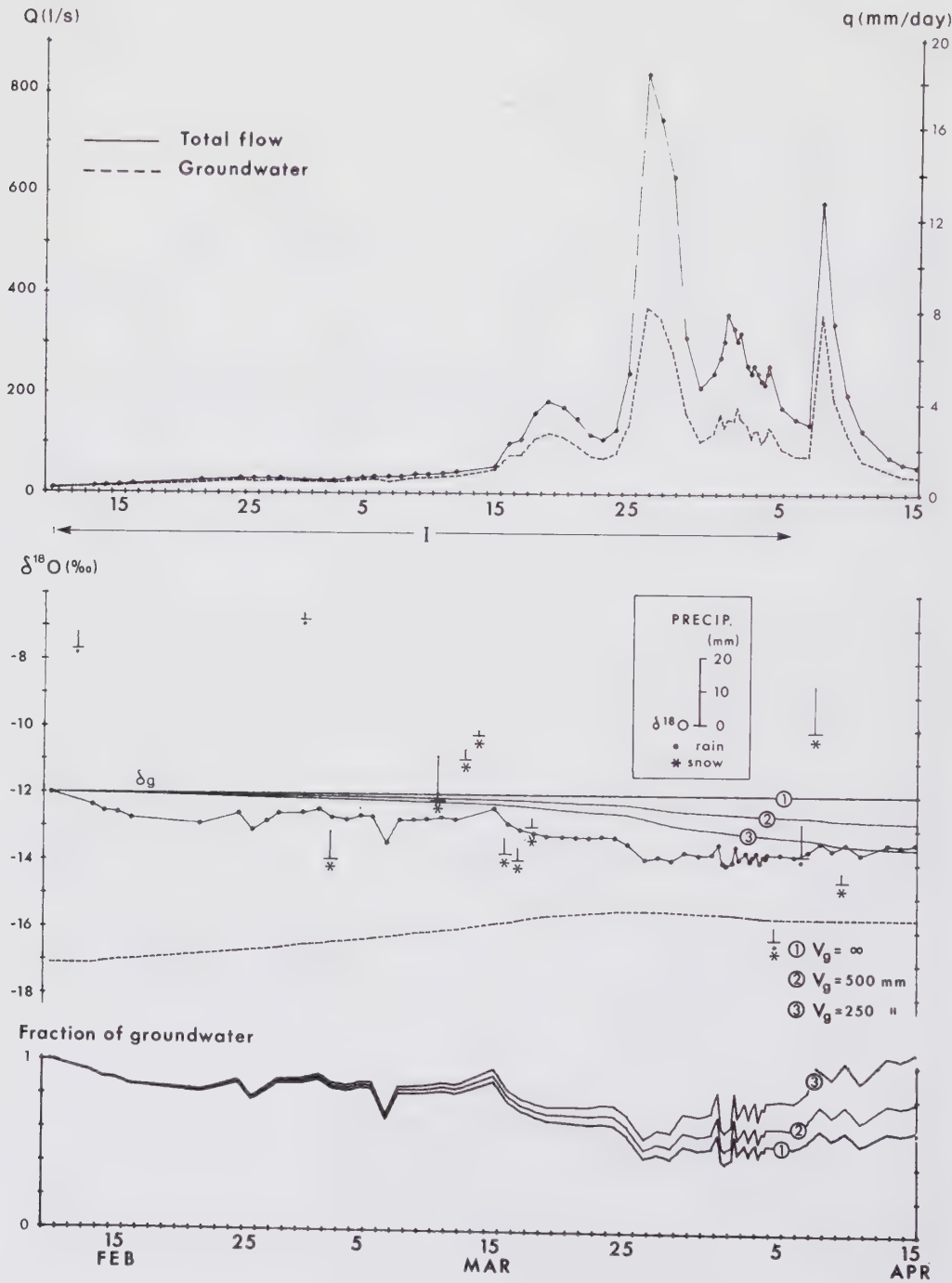


Fig.A2.16 Svartberget Nedre 1/4 - 29/5 1981. Flow of groundwater calculated for different values of V_g .

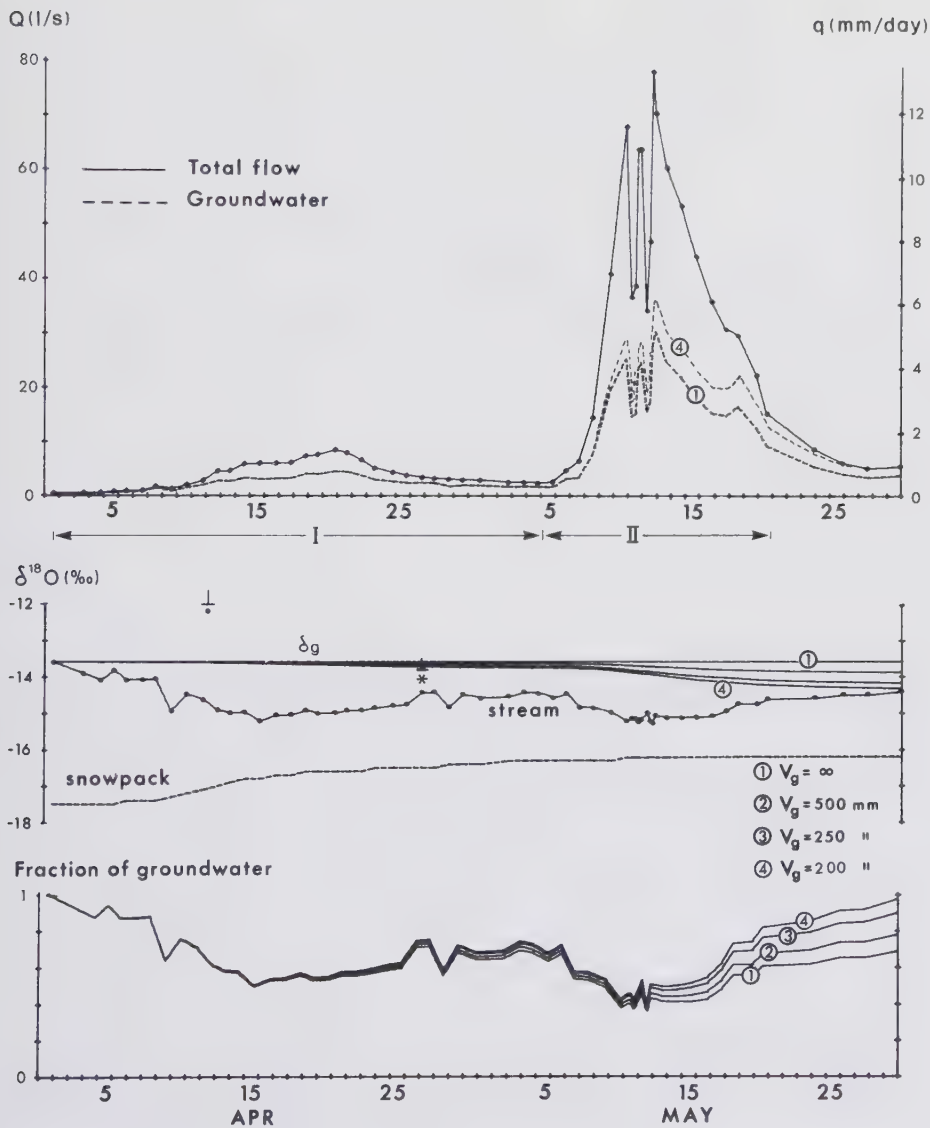


Fig.A2.17 Svartberget Nedre 25/3 - 22/5 1982.

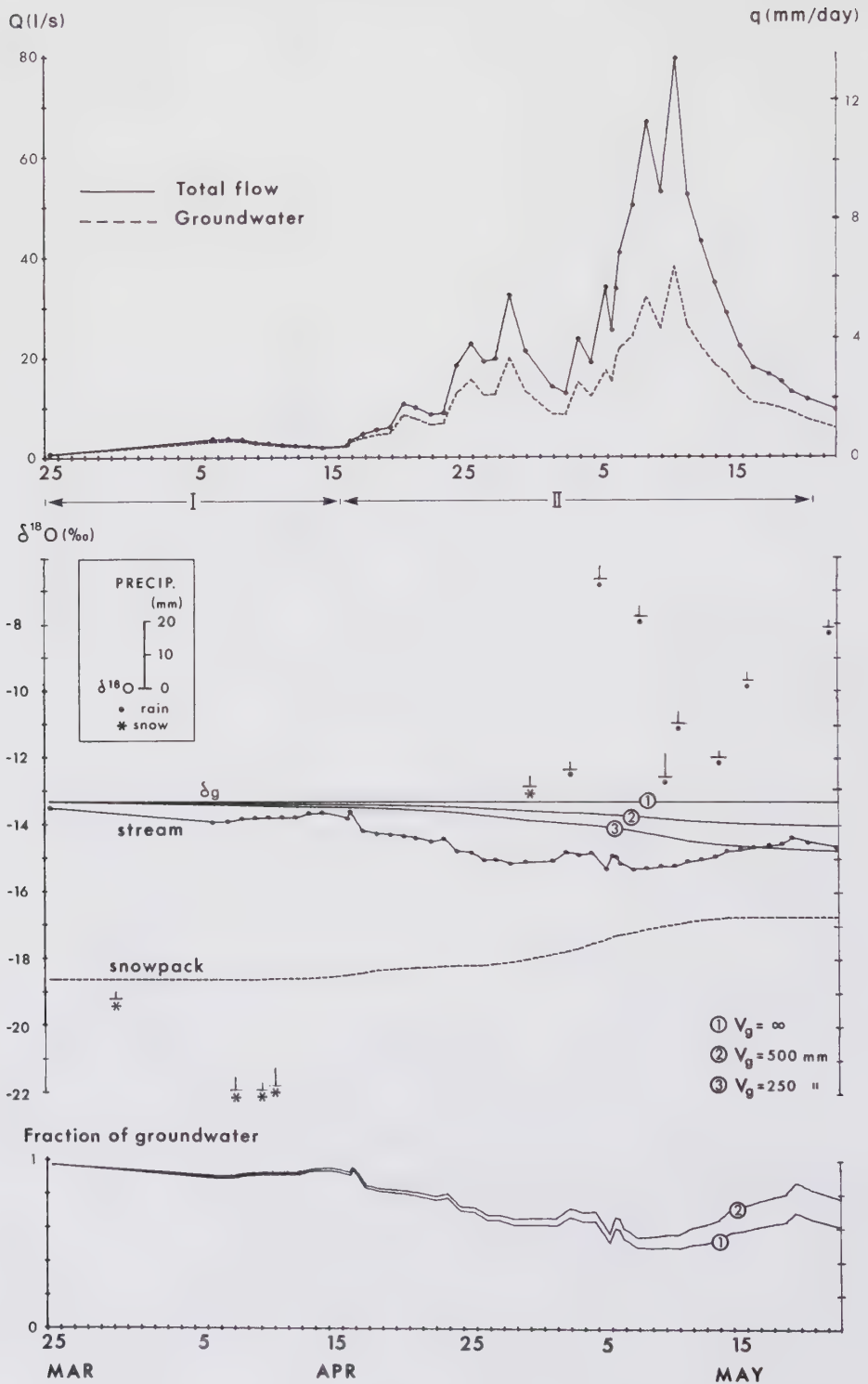


Fig.A2.18 Svartberget Övre (upper diagram) and Svartberget Västra (lower diagram) 1/4 - 21/5 1981. Stream hydrograph from Svartberget Nedre, with groundwater flow estimated from ^{18}O observations Svartberget Övre.

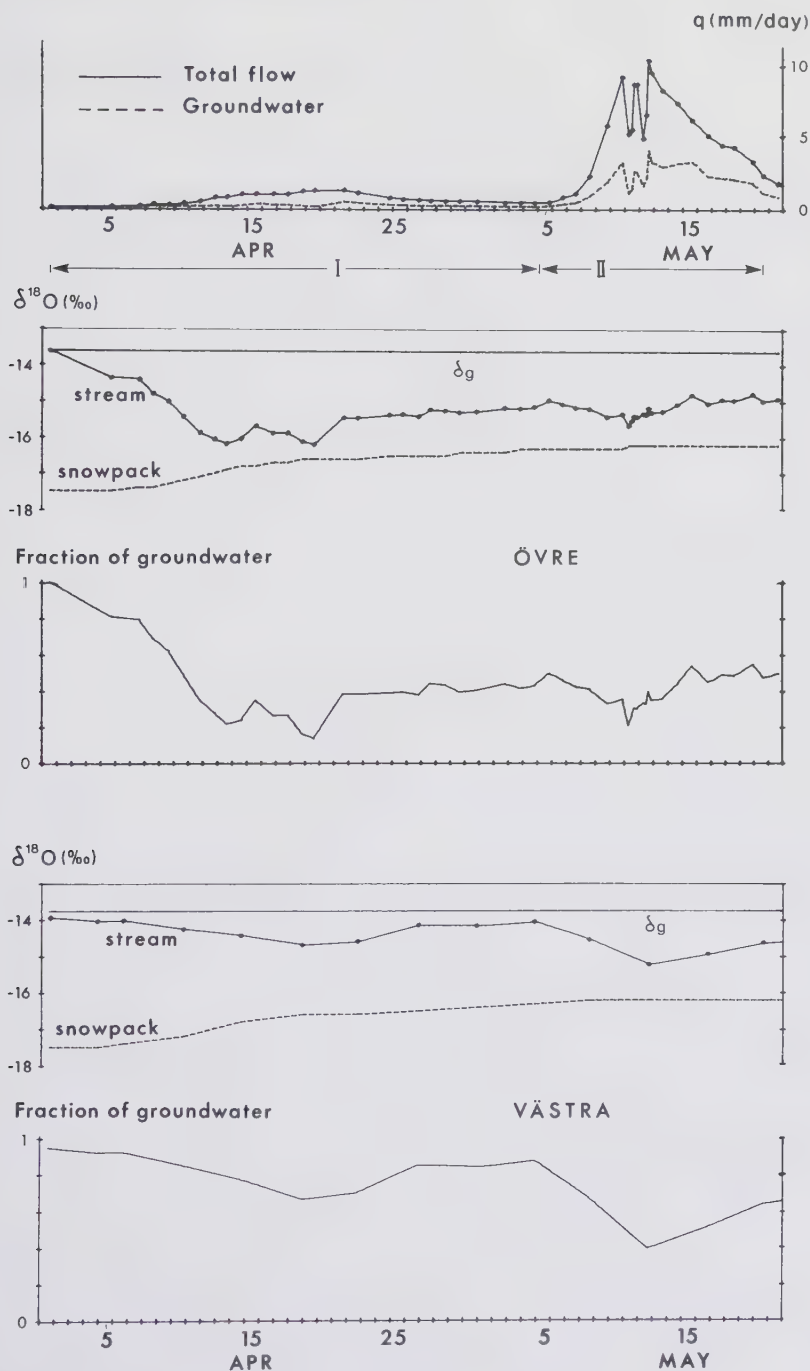
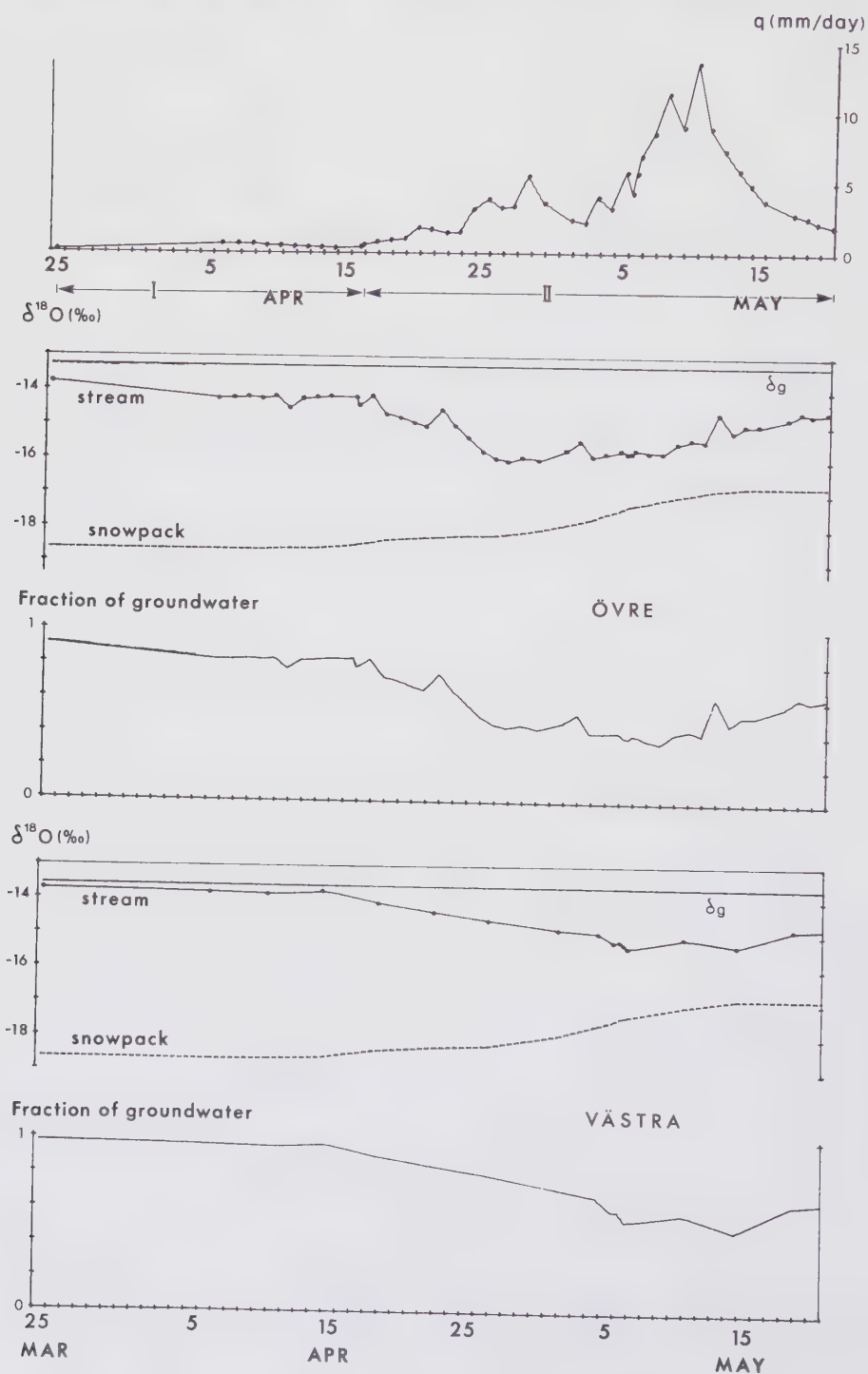


Fig.A2.19 Svartberget Övre (upper diagram) and Svartberget Västra (lower diagram) 25/3 - 20/5 1982. Stream hydrograph from Svartberget Nedre.



APPENDIX 3 RAINFALL EVENTS

Table on results of hydrograph separations and hydrologic conditions of the rainfall events

Rainfall events not presented in the main text, annotated figures

Table A3.1 Results of hydrograph separation: rainfall events

Basin			Period	Total rainfall (mm)	Maximum 15-min rainfall intensity (mm/h)	Maximum discharge (l s ⁻¹ km ⁻²)
Aspåsen	A	1985	19/9 -22/9	14	2	88
Buskbäcken	A	1980	23/6 - 1/7	75	-	220
Gårdsjön	F1 A	1982	12/6 -17/6	63	6	111
	F1 B	1982	3/7 - 6/7	17	16	11
	F3 A	1982	12/6 -17/6	63	6	90
	F3 B	1982	3/7 - 6/7	17	16	16
Nåsten	A	1979	23/11- 2/12	43	5	98
	B	1979-80	28/12-13/1	33	-	32
	C	1980	30/5 - 8/6	16	5	11
	D	1980	7/8 -15/8	49	3	15
	E	1982	3/7 -11/7	26	20	2
Stormyra	A	1979	25/8 - 1/9	47	15	19
	B	1979	15/10-19/10	22	-	6
	C	1980	1/6 - 9/6	47	54	145
	D I	1980	6/7 -11/7	17	50	11
	D II	1980	16/7 - 1/8	47	13	120
Svartberget	A	1981	17/8 -25/8	31	2	54
Nedre	B	1981	29/9 - 4/10	18	4	22
	C	1981	20/10- 2/11	40	-	30

¹ Runoff within the period² Runoff including recession³ Calculated by constant $\delta^{18}\text{O}$ of groundwater

Total runoff ¹ (mm)	Discharged fraction of rainfall ² (%)	Fraction of groundwater ³ (%)	Fraction of discharge area (%)
8	55	87	7
31	47	-	-
13.6	21	81	4
1.5	9	96	0.4
8.5	13	81	2.5
1.7	10	87	1.3
41	95	81	17
13	47	78	8
3.6	22	≈99	≈0
5.6	14	>66	<4
0.6	23	95	0.2
5.3	14	81	2.1
2.9	17	86	1.8
21.5	46	≈68	≈15
2.3	14	>63	<3.7
24.5	52	>53	<25
11.1	35	92	3.0
4.1	23	85	3.5
16.2	40	99	0.4

RAINFALL EVENTS NOT PRESENTED IN THE MAIN TEXT, ANNOTATED FIGURES

Aspäsén A, September 1985 (Fig A3.1)

Fig.A3.1 Stream hydrograph with calculated flow of groundwater, $\delta^{18}\text{O}$ of streamwater, and $\delta^{18}\text{O}$ and amount of rainfall (legend, see Fig. A3.2). Aspäsén A, September 1985.

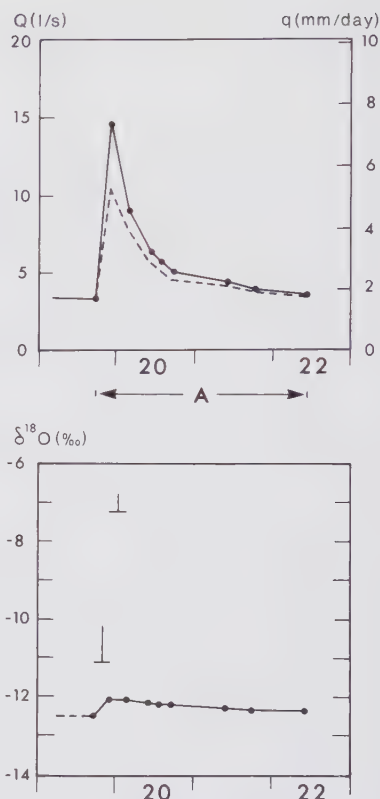
The runoff event was generated by a 14.4 mm rainfall having an intensity of around 2 mm h^{-1} . The accumulated weighted mean of the samples was used as δ_p . With the observed variation of rainfall $\delta^{18}\text{O}$, this procedure probably underestimates the total volume fraction of groundwater (Section 8.2.1).

As mentioned in Chapter 4, the basin was 86% clearing at this event, in contrast to the situation at the separated snowmelt event when the basin was completely vegetated by forest. The basin was already very wet at the onset of rainfall, with streamflow receding from an extremely high discharge of 108 l/s on September 4. But since the basin is comparatively steep and of simple shape, there was little water stored as surface water, only in some tractor tracks and in the stream itself.

The discharge areas were mapped during the event, giving 15 % of the basin area. This value is larger than the one obtained by Eq(3.9), 7%. One reason for the discrepancy might be that the mapped area, which included the whole lower part of the basin, contained several spots of unsaturated soil where rainwater could be stored.

Buskbäcken A, see main text, p. 189.

Gårdsjön F1 A and F1 B, see main text, pp. 179 and 182.



Gårdsjön F3 A and B, see main text p. 184.

Nåsten A, November 1979 (Fig.A3.2)

The first of the two large runoff events shown in Fig.A3.2 could not be separated, since the important rainfalls were either lighter or heavier in ^{18}O than the initial streamwater. The second event, referred to as Nåsten A, could be roughly separated, but there might be a considerable influence of pre-event surface water.

The accumulated weighted mean was taken as δ_p , starting on November 24. Since the daily values of rainfall $\delta^{18}\text{O}$ deviated progressively from that of the initial streamwater, this procedure tends to underestimate the groundwater fraction.

The $\delta^{18}\text{O}$ of streamwater during the first peak was taken as δ_g . This also tends to underestimate the groundwater fraction, but due to the large difference in the $\delta^{18}\text{O}$ of the flow components, the calculated groundwater fraction is not very sensitive to errors in δ_g .

The dominance of groundwater is clear from the isotopic data. The fact that streamwater $\delta^{18}\text{O}$ did not return to its initial value indicates that the $\delta^{18}\text{O}$ of groundwater contributing to streamflow changed during this large event. A model reservoir volume of 500 mm in Eq.(3.6) roughly describes the total change in groundwater $\delta^{18}\text{O}$, yielding a groundwater fraction of total discharged volume of 86% as compared to 81% by constant δ_g .

The large fraction of the rainfall being discharged during this event, estimated at 95%, is to be noted. It is the result of rainfall on a very wet basin during a season when potential evapotranspiration is very small.

Nåsten B, December 1979-January 1980 (Fig.A3.3)

A cold period, with air temperatures below freezing, ended on December 23. A snowpack, reaching a water equivalent around 6 mm, was formed during the period. Melting of the snowpack, in connection with some sleet, caused a small increase in discharge around December 24. Large rainfalls on December 28 and 29 generated a considerable discharge peak, which could be roughly separated by ^{18}O .

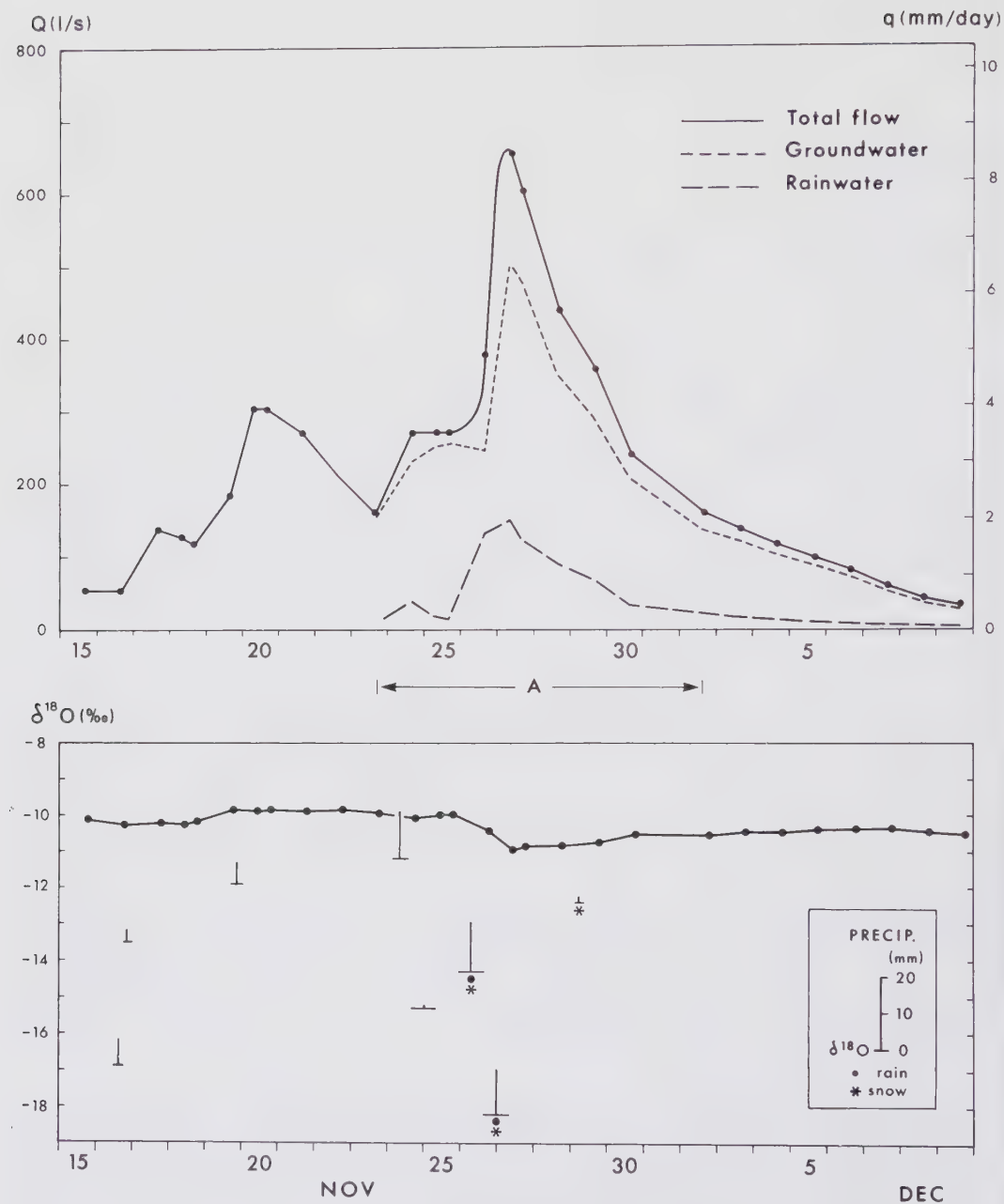


Fig.A3.2 Stream hydrograph with calculated flow of groundwater and rain- or meltwater, $\delta^{18}O$ of streamwater, and $\delta^{18}O$, amount and type of precipitation. The length of the horizontal precipitation lines shows the duration of precipitation when known; if the duration is not known the length marks the 24 h sampling period. Nâsten A, November 1979.

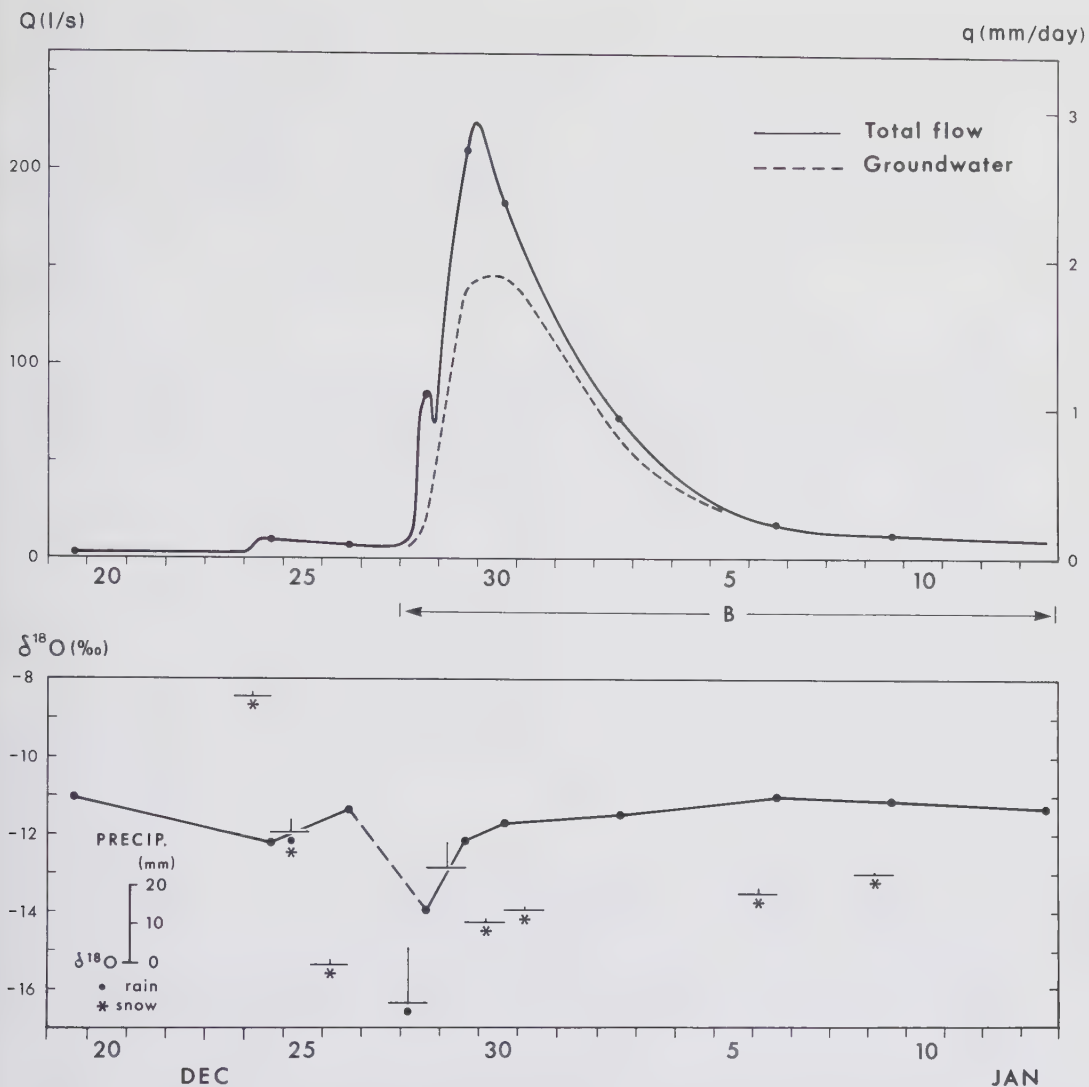


Fig.A3.3 Stream hydrograph with calculated flow of groundwater, $\delta^{18}O$ of streamwater, and $\delta^{18}O$, amount and type of precipitation. Nåsten B, December 1979 - January 1979.

The accumulated weighted mean $\delta^{18}O$ was taken as the $\delta^{18}O$ of rainwater (and meltwater), starting at the melting on December 23. It was hereby assumed that the meltwater amount and $\delta^{18}O$ were equal to those of the snow precipitation collected during the cold period. Since the rainfall on December 29 (6.6 mm, $\delta = -12.9\text{‰}$) was higher in ^{18}O than the preceding accumulated mean, the groundwater fraction may be overestimated as from this day.

The low value of streamwater $\delta^{18}\text{O}$ in the sampling of December 28, yielding a calculated instantaneous groundwater fraction of 0.25, is noteworthy. It agrees with the findings in some other events showing large relative contributions of rainwater during the hydrograph rise. A close inspection of the water-level recording chart shows a small extra peak around the moment of streamwater sampling on December 28, as marked in the hydrograph.

Nåsten C, June 1980 (Fig.A3.4)

The isotopic conditions for separation were good at this summer event, since all rainfalls associated with the discharge peak were substantially higher in ^{18}O than the initial streamwater.

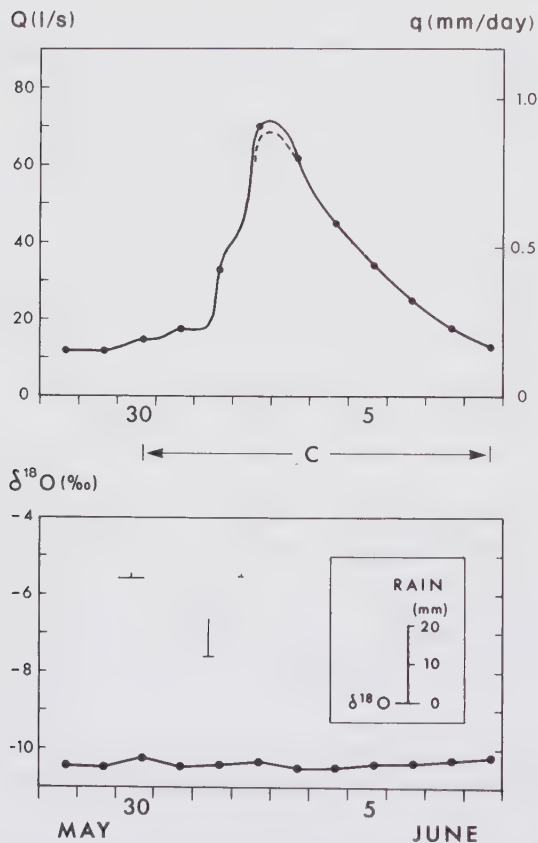


Fig.A3.4 Stream hydrograph with calculated flow of groundwater, $\delta^{18}\text{O}$ of streamwater, and $\delta^{18}\text{O}$ and amount of rainfall, (legend, see Fig.A3.2). Nåsten C, May - June 1980.

Surprisingly, no significant change in streamwater $\delta^{18}\text{O}$ was seen during the peak. Thus the groundwater flow, calculated using the weighted mean of rainfall $\delta^{18}\text{O}$ from June 1 and 2, is almost identical with the total flow and cannot be distinguished in

Fig.A3.4. The separation is not very sensitive to errors in δ_g or δ_p , so errors in the assessment of these variables are not likely to explain the apparent lack of direct rainwater contribution to streamflow. Denser sampling during the time of hydrograph rise might have revealed some rainwater contribution and slightly changed the groundwater hydrograph. But since two samples were taken near the time of peakflow, and since isotopic fluctuations during recession are very unlikely, the shortage of data on streamwater $\delta^{18}\text{O}$ cannot seriously have affected the calculated groundwater fraction of total flow volume.

As in the other events, the electrical conductivity of streamwater showed a dilution inversely related to stream discharge (from 38 to 26 mS/m). But due to the large variations of the observed electrical conductivity of groundwater in the basin, from 10 to 70 mS/m, this tracer cannot be used for two-component hydrograph separation.

Nåsten D, August 1980 (Fig.A3.5)

Except for a small rainfall (3.3 mm, $\delta = -7.5\text{‰}$), on August 7) all rainfalls during the hydrograph rise were lower in ^{18}O than the initial streamwater. During recession, on August 16, an isotopically heavy rainfall occurred, preventing separation from that day onwards.

As in Nåsten C, the streamwater samples show little influence of rainwater on the isotopic composition of streamwater. The initial $\delta^{18}\text{O}$ of streamwater is somewhat uncertain, since values of streamwater $\delta^{18}\text{O}$ of $-10.4 - -10.8\text{‰}$ were noted from August 3 to 6. The separation was made up to August 15 using $\delta_g = -10.4\text{‰}$ and weighted 2-day means as δ_p .

Due to the sparse streamwater sampling the total volume fraction of groundwater cannot be given. But considering that a decrease in groundwater fraction after peakflow is unlikely, the groundwater hydrograph is extrapolated, establishing a rough lower limit of groundwater fraction up to August 15 of 66%.

Nåsten E, I and II, June-July 1982 (Fig.A3.6)

The isotopic conditions for separation of these two, very small, events were acceptable. But during the first event, rainwater was lower in ^{18}O than the initial streamwater and during the second event, it was higher.

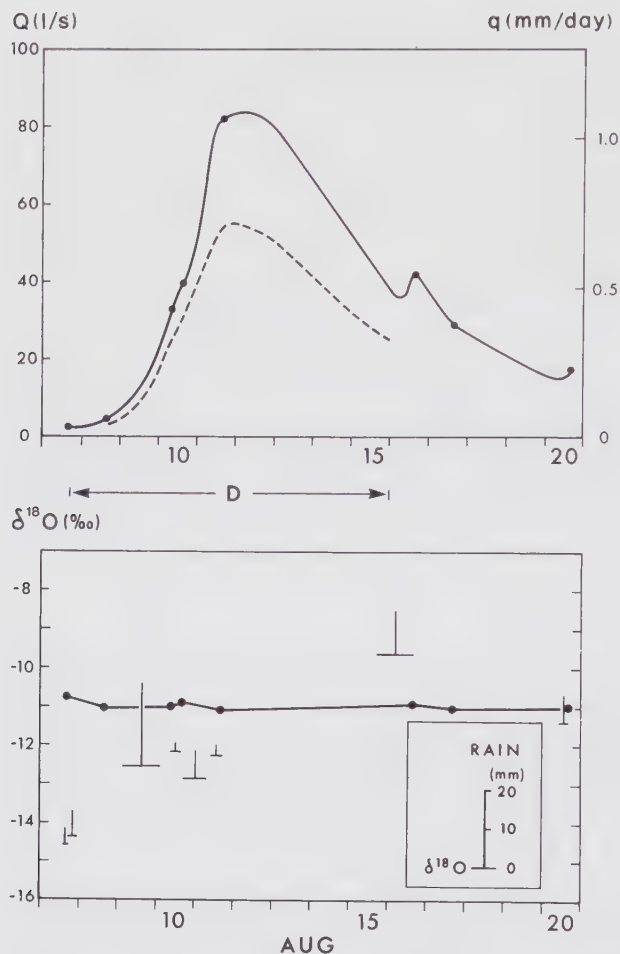


Fig.A3.5 Stream hydrograph with calculated flow of groundwater, $\delta^{18}\text{O}$ of streamwater, and $\delta^{18}\text{O}$ and amount of rainfall, (legend, see Fig. A3.2). Nåsten D, August 1980.

Different values of δ_g were used for the two events, -10.0 and -10.4 ‰, respectively. The first value is that of pre-event streamwater, while the second one is streamwater $\delta^{18}\text{O}$ during the first event. By the latter choice a lower limit of groundwater fraction is obtained. The weighted mean $\delta^{18}\text{O}$ of the first two rainfall samples was used as δ_p for event I. For event II rainfall $\delta^{18}\text{O}$ of July 3 was used. Due to the small difference between δ_p and δ_g , the separation of event I is uncertain. The results may also be affected by enrichment during surface water flow at these very low flows in summertime. For this reason, event I is not listed in Table 8.6 and not included in the general discussions.

Stormyra A, see main text, p. 185.

Stormyra B, see main text, p. 185.

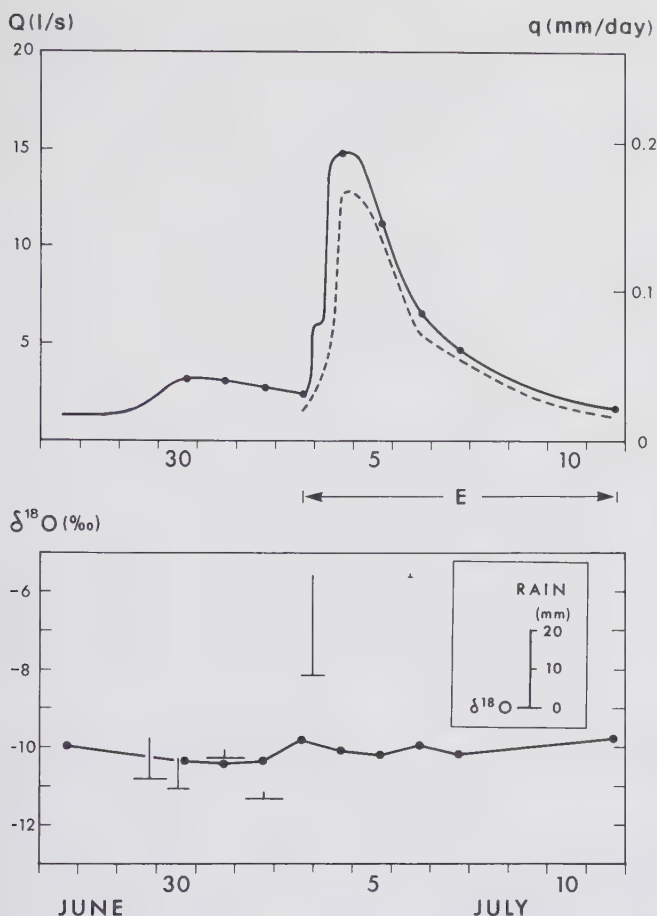


Fig.A3.6 Stream hydrograph with calculated flow of groundwater, $\delta^{18}\text{O}$ of streamwater, and $\delta^{18}\text{O}$ and amount of rainfall, (legend, see Fig. A3.2). Nåsten E, June - July 1982.

Stormyra C, June 1980 (Fig.A3.7)

A heavy thunderstorm delivered a rainfall of 47 mm in the late evening of June 1. The rainwater was considerably higher in ^{18}O than the initial streamwater. Streamwater sampling of this large event was very sparse, but fortunately one of the samples was taken at the moment of peakflow. The minor rainfalls on May 30 and 31 were disregarded in the separation.

According to isotope data, the instantaneous groundwater fraction was 0.60 at peakflow. In the recession samples it was 0.80 and 0.85, respectively. Although the streamwater sampling is very sparse, the volume fraction of groundwater can be acceptably estimated. This is so since the groundwater fraction is unlikely to increase after peakflow and since a little fraction

of the total discharge appears before peakflow. The groundwater fraction is estimated at 68% of the total discharged volume. The shapes of the rainwater and groundwater hydrographs, on the other hand, cannot be obtained. Smaller groundwater fractions are likely to have occurred during the hydrograph rise.

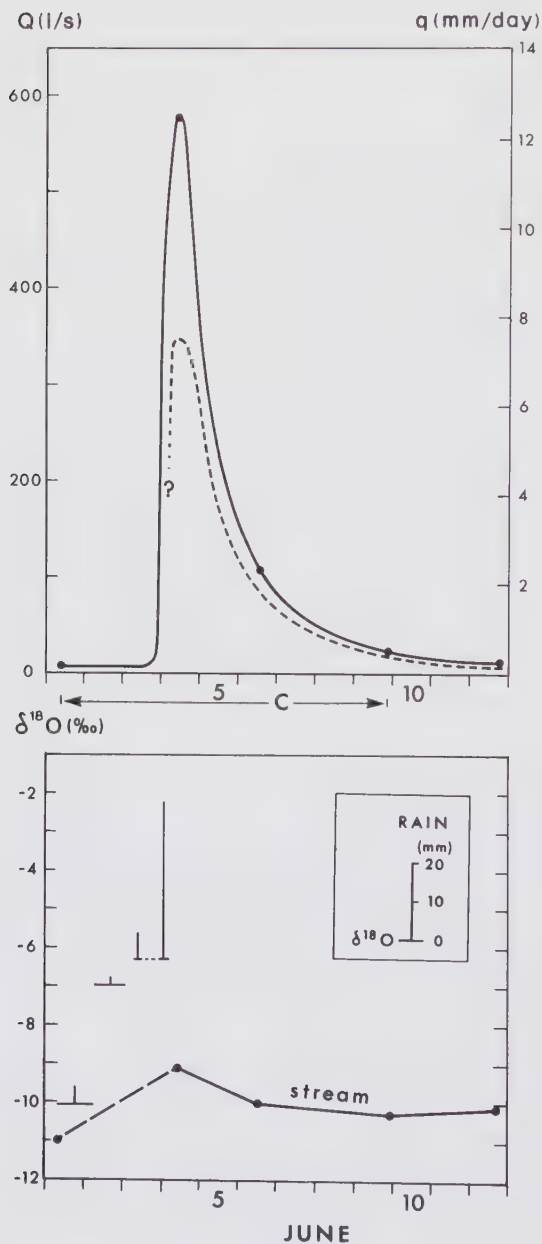


Fig.A3.7 Stream hydrograph with calculated flow of groundwater, $\delta^{18}\text{O}$ of streamwater, and $\delta^{18}\text{O}$ and amount of rainfall, (legend, see Fig.A3.2). Stormyra C, June 1980.

A model reservoir in Eq.(3.6) of 100 mm roughly describes the observed net change in streamwater $\delta^{18}\text{O}$ during the event, i.e., the indicated change in groundwater $\delta^{18}\text{O}$.

Stormyra D, I and II, July 1980 (Fig.A3.8)

The isotopic and hydrologic conditions for hydrograph separation were good at these two events until the rainfall on July 20, which, in contrast to the earlier rainfall, was higher in ^{18}O than the initial streamwater. For the first event the total groundwater fraction could be roughly estimated. But since there are no samples from peakflow of the second event, the sparse streamwater sampling does not allow the total fraction of groundwater to be estimated for this one. Lower limits of the fraction can, however, be established.

Accumulated weighted means of rainfall $\delta^{18}\text{O}$, calculated for each event, were used in the separations. The $\delta^{18}\text{O}$ of streamwater on July 1, which was before any rainfall occurred, was taken as δ_g for both events.

At the large event, one streamwater sample was taken during the hydrograph rise. This sample showed the smallest instantaneous fraction of groundwater during the events, 54%.

Estimated lower limits of total groundwater fractions, according to the groundwater hydrographs sketched in Fig.A3.8, were 63% and 53%, respectively, for the events. (The second value refers to the total fraction up to June 20, when the isotopically unfavourable rainfall occurred). The limit given for the large event is conservative, since streamwater $\delta^{18}\text{O}$ is unlikely to change rapidly during recession, which was presumed when sketching the groundwater hydrograph around peakflow.

Svartberget Nedre A, see main text, p. 191.

Svartberget Nedre B, September-October 1981 (Fig.A3.9)

In contrast to Svartberget Nedre A, this event was generated by rainfalls higher in ^{18}O than the initial streamwater, and the observed increase in streamwater $\delta^{18}\text{O}$ was a change in the expected direction. The $\delta^{18}\text{O}$ of the different rainwater samples show little scattering until October 4, after which day separation could not be performed. Using the initial streamwater

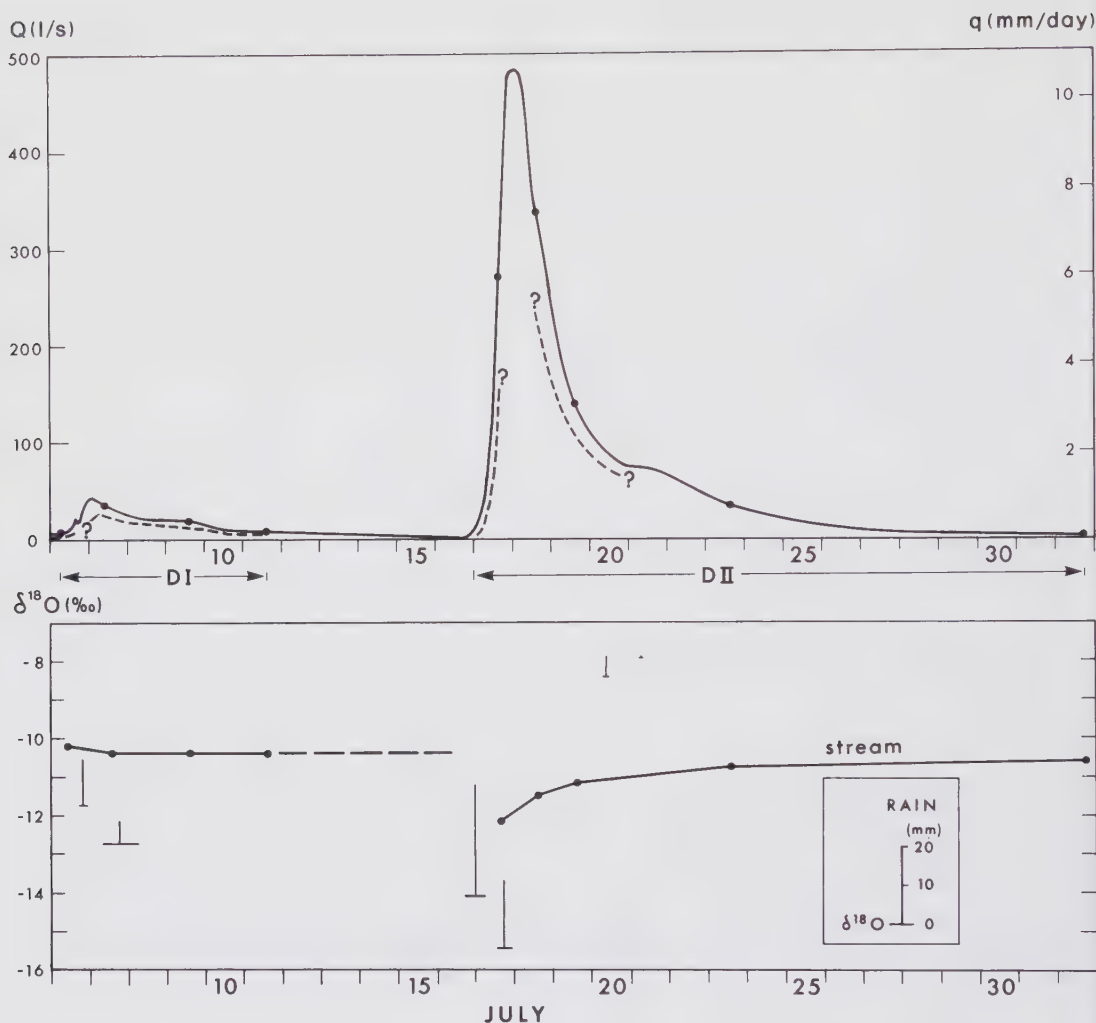


Fig.A3.8 Stream hydrograph with calculated flow of groundwater, $\delta^{18}O$ of streamwater, and $\delta^{18}O$ and amount of rainfall, (legend, see Fig.A3.2). Stormyra D I and II, July 1980.

$\delta^{18}O$ as δ_g and the $\delta^{18}O$ of the main rainfall as δ_p , the total volume fraction of groundwater up to October was 85%. There were no streamwater samples taken during the hydrograph rise, but possible deviations of streamwater $\delta^{18}O$ during the rise would make only minor changes in the calculated total fraction.

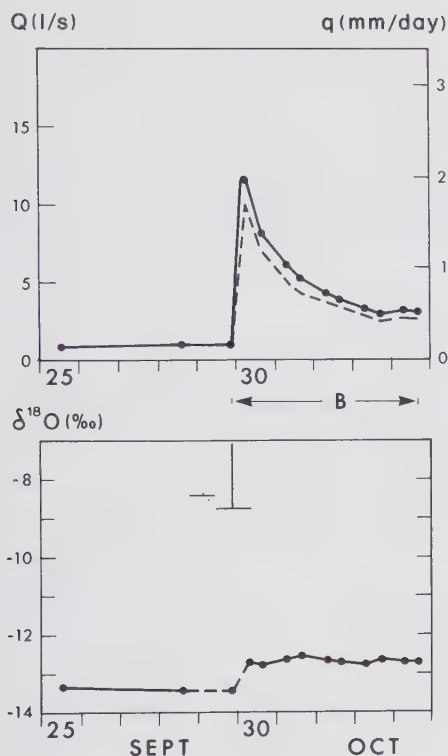


Fig.A3.9 Stream hydrograph with calculated flow of groundwater, $\delta^{18}\text{O}$ of streamwater, and $\delta^{18}\text{O}$ and amount of rainfall, (legend, see Fig.A3.2). Svartberget Nedre B, September - October 1981.

Due to the large difference between δ_p and δ_g , the separation is not very sensitive to errors in δ_g , which may have been changing during the event (see the discussion on Svartberget Nedre A).

Svartberget Nedre C, October 1981 (Fig.A3.10)

The rainfall and snowfall during this event were very low in ^{18}O as compared to the initial streamwater; still the changes in streamwater $\delta^{18}\text{O}$ were small, indicating little direct rainwater contribution. The $\delta^{18}\text{O}$ of the rainfall on October 20 was used in the separation throughout the event. Possible influence of the very ^{18}O -depleted snowfall on October 21 would give still larger groundwater fractions.

The estimated groundwater hydrograph is almost identical with the total hydrograph and gives a total fraction of groundwater

of 99%, comparatively insensitive to errors in δ_g and δ_p . The smallest instantaneous fraction of groundwater, 93%, was encountered during the hydrograph rise.

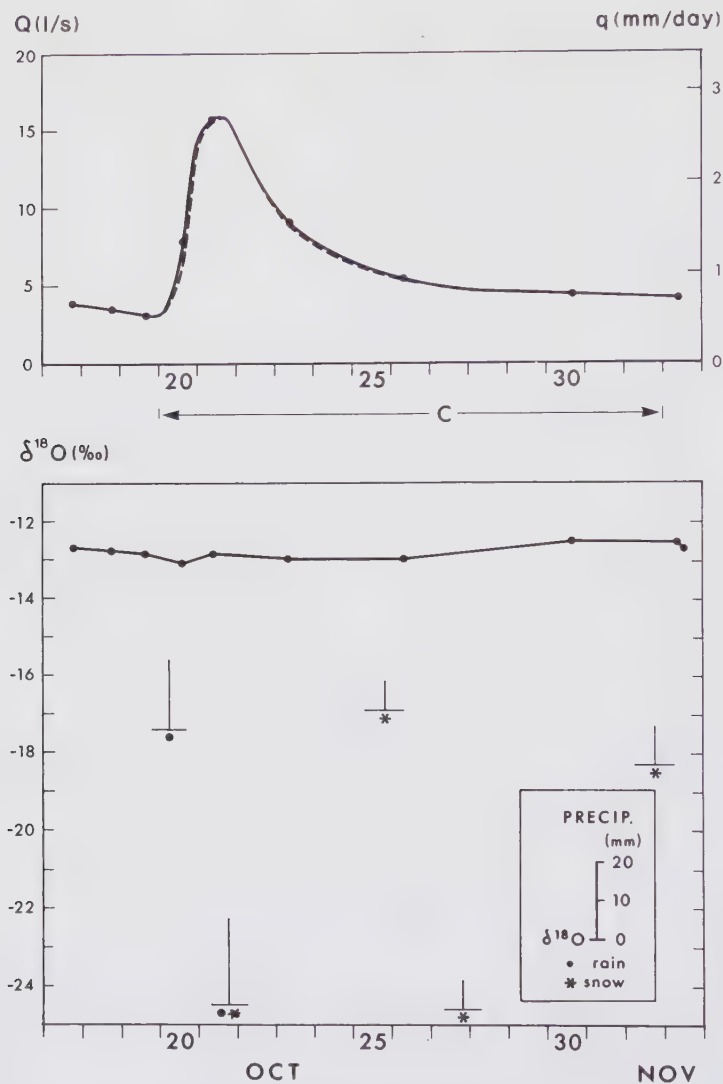


Fig.A3.10 Stream hydrograph with calculated flow of groundwater, $\delta^{18}\text{O}$ of streamwater, and $\delta^{18}\text{O}$ and amount of rainfall, (legend, see Fig.A3.2). Svartberget Nedre C, October 1981.

APPENDIX 4 INFLUENCE OF PRECIPITATION ON HYDROGRAPH
SEPARATIONS DURING SNOWMELT

Influence of precipitation on hydrograph separations during snowmelt

With the considerations in Section 9.2, a corrected value of X , X_{corr} , can be obtained from mass and isotope balance for the four flow components: groundwater (G , δ_g), meltwater (M , δ_p), melted snowfall (P'_s , δ_{ps}), and rainwater (P'_r , δ_{pr}), mixing into streamflow (Q , δ_s):

$$G\delta_g + M\delta_p + P'_s \delta_{ps} + P'_r \delta_{pr} = Q\delta_s \quad (\text{A4.1})$$

$$G + M + P'_s + P'_r = Q \quad (\text{A4.2})$$

giving

$$\frac{G}{Q} = \frac{\delta_s - \delta_p}{\delta_g - \delta_p} - \frac{P'_s(\delta_{ps} - \delta_p) + P'_r(\delta_{pr} - \delta_p)}{Q(\delta_g - \delta_p)} \quad (\text{A4.3})$$

or

$$X_{\text{corr}} = X - \frac{P'_s(\delta_{ps} - \delta_p) + P'_r(\delta_{pr} - \delta_p)}{Q(\delta_g - \delta_p)} \quad (\text{A4.4})$$

The application of Eq.(A4.4) is equivalent to the application of Eq.(3.3) using a corrected value of δ_p , $\delta_{p \text{ corr}}$, defined as

$$\delta_{p \text{ corr}} = \delta_p - \frac{1}{Q} [P'_s(\delta_{ps} - \delta_p) + P'_r(\delta_{pr} - \delta_p)] \quad (\text{A4.5})$$

If the flows of water are expressed per basin area, the direct contributions from snowfall, P'_s , and rainfall P'_r , are given by

$$P'_s = Y'P_s \quad (\text{A4.6})$$

$$P'_r = Y''P_r \quad (\text{A4.7})$$

where P_s and P_r are the total snowfall and rainfall respectively, while Y' and Y'' are discharged fractions of the new snowfall and rainfall over the basin. Y' is then assumed to be the basin fraction of snow-free discharge area, since only snowfall on such areas is likely to reach the stream directly. Y'' depends on the rainwater storage in the snowpack and on the disposition of the rainfall to generate overland flow. The effect of isotopic exchange between percolating rainwater and snowpack, tending to decrease $\delta_{pr} - \delta_p$ for $\delta_{pr} > \delta_p$, can be accounted for in Eq.(A4.7) by a corresponding relative decrease

in P_r . According to the hypothesis for the present study, $Y'' < Y$, where Y is the basin fraction of discharge area; but this, of course, cannot be postulated.

Due to water storage in the channel network, the correction cannot be applied to daily values of precipitation. An idea of the influence can be obtained from period mean values, however.

The influence of precipitation, as expressed by Eq.(A4.4), has been estimated for each snowmelt period using period means of precipitation, snowpack, and stream data (Table 9.2). Two values of $\Delta X = X_{\text{corr}} - X$ have been calculated, using different assumptions on Y'' :

$$\Delta X_1 \text{ from } Y' = 0.5 Y \text{ and } Y'' = 0.5 Y$$

$$\Delta X_2 \text{ from } Y' = 0.5 Y \text{ and } Y'' = 2 Y$$

With the latter assumptions, possibilities of Hortonian overland flow of rainwater from outside the saturated areas are taken into account.

The estimates of Y' and Y'' are based on the values of Y , determined by uncorrected values of δ_p . An iteration procedure would be more correct, with Y being successively determined from corrected values of δ_p . But since the steps from Y to Y' and Y'' are mere guesses, only the uncorrected values of Y are used.

APPENDIX 5 MOLECULAR EXCHANGE BETWEEN OVERLAND FLOW
AND GROUNDWATER

Molecular exchange between overland flow and groundwater

The isotopic composition of water flowing as overland flow may be modified by molecular exchange with water in the soil on which it flows. A necessary condition for such a modification is that the vertical diffusive flow of ^{18}O through the soil surface be larger than the advective flow, i.e., the flow caused by water flow. The exchange may occur during downward as well as upward water flow, provided that the diffusive flow dominates (cf. discussion on enrichment by fractionation at the surface of a non-mixed body of water). The largest modification of the surface water is to be expected when the groundwater table reaches the soil surface, but no outflow or inflow of groundwater occurs. The isotopic changes of a water film flowing over such an area is estimated below. The problem, which never seems to have been dealt with in the literature, will be treated in some detail. A general expression for the isotopic change is deduced, and attempts are made to apply it to situations met in the present study.

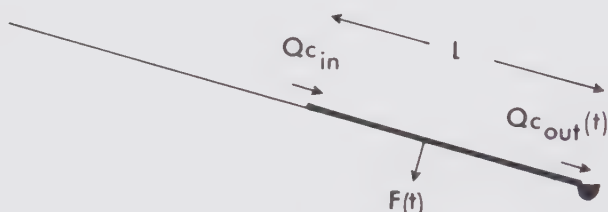


Fig.A5.1 Tracer balance of water flowing on a saturated area. Q = rate of overland flow, c = tracer concentration, l = length of the saturated area, F = diffusive tracer flux to the groundwater.

Consider the situation shown in Fig.A5.1. A saturated area with no groundwater flow through the surface extends a distance l up the slope from the stream. Overland flow, having an original tracer concentration c_{in} , passes the saturated area at a rate Q per width unit. For stationary water flow, the tracer concentration of the water leaving the saturated area at a time t , $c_{out}(t)$, is obtained by tracer balance

$$Qc_{in} - Qc_{out}(t) - lF(t) = 0 \quad (\text{A5.1})$$

where $F(t)$ is the flux of tracer to the groundwater. (The storage of tracer in the surface water is disregarded, making the right hand side of Eq.A5.1 = 0.) Note that t is the time

from the beginning of the flow situation, not the time of travel over the saturated area. The tracer change, $\Delta c(t) = c_{out}(t) - c_{in}$ is given by Eq.(A5.1) as

$$\Delta c(t) = - \frac{1}{\frac{Q}{1}} F(t) \quad (A5.2)$$

The diffusive flux, F , is estimated using the theory of molecular diffusion in solids. To simplify the problem we assume that the tracer concentration at the ground surface is constantly kept at $c = c_{in}$ (Fig.A5.2). In this way an upper limit of F is obtained, since in reality $c - c_g$ decreases downslope, reducing the diffusion of tracer into the groundwater.

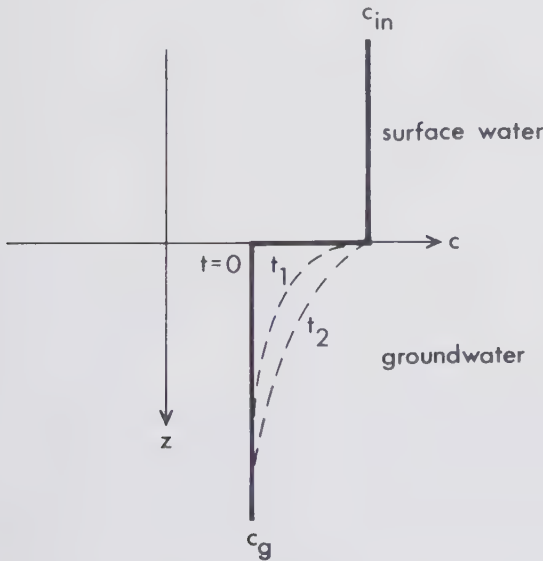


Fig.A5.2 Vertical distribution of tracer concentration in surface water and groundwater. c_{in} and c_g are tracer concentrations in surface water and groundwater respectively, z depth, and t time.

In analogy with the one-dimensional distribution of temperature in a semi-infinite solid originally having a constant temperature, but with the surface kept at another temperature from time $t = 0$ (see Carslaw and Jaeger 1962, p.58), we get the vertical tracer distribution in groundwater as

$$c(z,t) - c_g = (c_{in} - c_g) \left(1 - \operatorname{erf} \frac{z}{2 \sqrt{Dt}} \right) \quad (A5.3)$$

where $c(z,t)$ = tracer concentration of groundwater
 c_g = original tracer concentration of groundwater
 c_{in} = tracer concentration of surface water
 z = depth

D = coefficient of molecular diffusion for the tracer in water

erf = error function; $\text{erf } x = \frac{2}{\sqrt{\pi}} \int_0^x e^{-\xi^2} d\xi$

The flux at the surface, $z = 0$, is given by

$$F(t) = -p D \left(\frac{\delta c(0, t)}{\delta z} \right) \quad (\text{A5.4})$$

where p is the porosity¹ of the soil. Now $\frac{\delta c(0, t)}{\delta z}$ is simply $-(c_0 - c_g)/\sqrt{\pi D t}$ (Carslaw and Jaeger, p.61) giving

$$F(t) = p D \frac{c_0 - c_g}{\sqrt{\pi D t}} \quad (\text{A5.5})$$

Defining the relative tracer change, $r(t)$, as

$$r(t) = \frac{\Delta c(t)}{c_{in} - c_g} \quad (\text{A5.6})$$

we get by Eqs.(A5.2) and (A5.5)

$$r(t) = - \frac{p}{\sqrt{\pi}} \frac{\sqrt{D}}{\frac{Q}{I} \sqrt{t}} \quad (\text{A5.7})$$

The largest tracer change is thus obtained at the beginning of situations with small flow rates. As the flow situation proceeds, the tracer gradient into the groundwater - and thereby the diffusive flux - decreases, making the tracer change of the overland-flowing water decrease.

By integration of Eq.(A5.7) we get the mean relative tracer change from time $t = 0$ (beginning of the flow situation) to time t ,

$$\bar{r}(t) = - \frac{2p}{\sqrt{\pi}} \frac{\sqrt{D}}{\frac{Q}{I} \sqrt{t}} \quad (\text{A5.8})$$

¹ The porosity does not appear in Eq.(A5.3) since the smaller cross-section area for diffusive flux in soil water as compared to "massive water" is compensated for by the smaller tracer storage capacity per bulk volume of the soil. The porosity therefore disappears in the mass balance equation to which Eq.(A5.3) is a solution.

The mean tracer change up to a certain moment is thus twice the instantaneous tracer change at the same moment,

$$\bar{r}(t) = 2r(t) \quad (\text{A5.9})$$

Looking at the whole basin, the term Q/l can be written q/Y'' , where q is the rate of overland flow per basin area and Y'' the basin fraction of saturated area with no groundwater flow through the surface. With $p = 0.5$ and $D = 1.68 \cdot 10^{-9} \text{ m}^2 \text{ s}^{-1}$ ($D \text{ H}_2^{18}\text{O}$ in water at $+10^\circ\text{C}$; Wang, 1965¹), Eq.(A5.7) becomes, with the corresponding δ -values

$$\bar{r} = \frac{\Delta\delta}{\delta_p - \delta_g} = \frac{1.4}{\frac{q}{Y''} \sqrt{t}} \quad (\text{A5.10})$$

for q in mm h^{-1} and t in h .

Possible values of q and Y'' during runoff production in the basins investigated and the resulting values of q/Y'' are given in Table A5.1.

The values of Y'' , 0.01 and 0.05, are mere guesses. The only related field data are those from surveys of saturated areas carried out during some of the spring floods. These values range from 0.10 to 0.35 of the basin areas. According to field experience, the fractions are considerably smaller during summer. But the extent to which saturated areas without groundwater outflow exist is not known.

In assessing values of q , two extremes were chosen. The large values in Table A5.1, q_1 , presume that all rainfall or snowmelt in the basin generates overland flow, giving $q_1 = i$, where i is the intensity of water input. The smaller values, q_2 , are obtained if overland flow is assumed to be generated on saturated areas only, giving $q_2 = i Y$, where Y is basin fraction of discharge area. In calculating q_2 it is assumed that $Y = 5 Y''$.

¹ According to Zimmermann et al. (1967), the molecular diffusion coefficient for water in soil is about 80% of its value in massive water. Considering the large variation of D with temperature (an increase by about $3\% \text{ } ^\circ\text{C}^{-1}$ around 10°C , Wang, 1965) and the crude assumption made on the porosity, this reduction is not taken into account here.

Table A5.1 Flow situations for which tracer changes by molecular exchange between overland-flowing water and groundwater are estimated.

Situation	Intensity of water input (mm h ⁻¹)	Y''	q ₁	q ₂ (mm h ⁻¹)	q ₁ /Y''	q ₂ /Y''	Dura- tion (h)
Intense rainfall	20	0.05	20	5	400	100	1
Medium rainfall	4	0.01	4	0.2	400	20	10
Intense snowmelt	2	0.05	2	0.5	40	10	100
Slow snowmelt	0.2	0.01	0.2	0.01	20	1	100

Y'' = basin fraction saturated area with no groundwater outflow

q₁ = maximum rate of overland flow (see text)

q₂ = minimum rate of overland flow (see text)

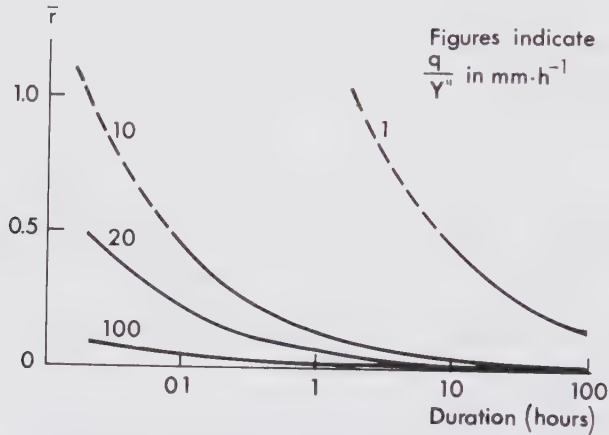


Fig.A5.3 Mean relative tracer change, $\bar{F}$, versus duration of flow for various flow rates. q is rate of overland flow per basin area and Y'' basin fraction of saturated area without groundwater outflow.

From Fig.A5.3 and Table A5.1 it is seen that the isotopic changes of the overland flowing water can be disregarded ($\bar{F} < 0.1$) during rainfall events, at least during those of large rainfall intensity. But snowmelt situations can be imagined in which the isotope changes according to Eq.(A5.10) cannot be disregarded. And even if the mean isotopic changes are small, the modification might be large during the first hours of low-intensity events. The conditions for diffusive modification are also seen in

Fig.A5.4, where the necessary duration (t) of different flow conditions (q/Y'') for the relative tracer change ($\bar{F}$) to fall below certain limits is shown.

In Fig.A5.3 the relative tracer change exceeds 1 for small values of t . With small values of q/Y'' , 1 is exceeded for a considerable time. This preposterous result stems from the assumption of constant tracer concentration of the surface water in the calculation of F . When $\bar{F} > 1$, the tracer inflow is smaller than the calculated diffusive flux. As pointed out earlier, the assumption results in an overestimation of the tracer change. But, except at the very smallest flow rates ($q/Y'' < 1 \text{ mm h}^{-1}$), the overestimation of the mean change is moderate for t of a few hours or more, since the lower start value of the true $\bar{F}$ - t curves is compensated for by delayed recession of the curves.

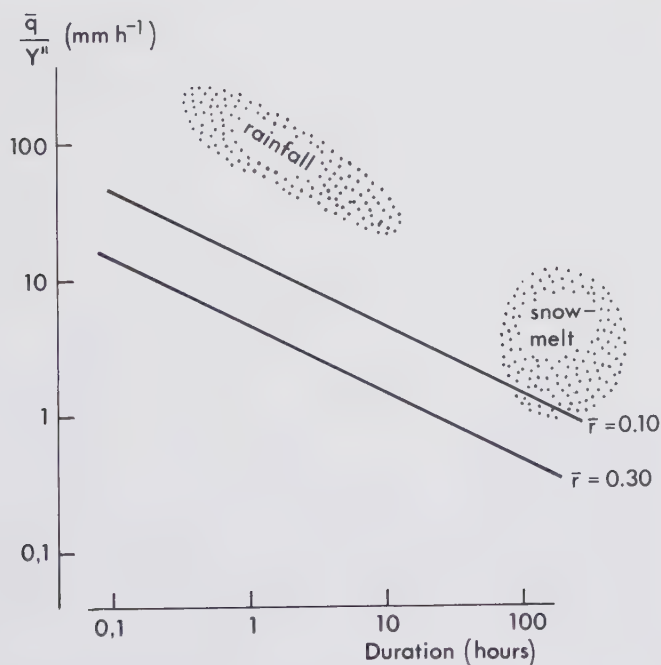


Fig.A5.4 Duration needed for various flow rates for the mean relative change, $\bar{F}$, to fall below certain values. q is rate of overland flow per basin area and Y'' basin fraction of saturated area without groundwater outflow. Dotted areas symbolize flow conditions to be expected for rainfall and snowmelt events respectively.

Eq.(A5.10) describes the effect of molecular diffusion in saturated areas with no groundwater outflow. The diffusive tracer changes of overland flowing water are largest in such areas. But

there may be (minor) diffusive changes also by overland flow on areas with water flow through the surface, downward (infiltration) as well as upward (groundwater outflow). These changes take place only at small vertical flow rates, when the diffusive flux dominates over the advective flow. The rate of groundwater outflow needed to prevent tracer changes of the surface water is estimated below.

The depth of tracer penetration, defined here as the depth $z_{0.1}$ at which $c(z,t) - c_g = 0.1(c_{in} - c_g)$, is obtained from Eq.(A5.3)

$$0.1 = 1 - \operatorname{erf} \frac{z_{0.1}}{2 \sqrt{Dt}} \quad (\text{A5.11})$$

giving

$$z_{0.1} = 2.3 \sqrt{Dt} \quad (\text{A5.12})$$

The mean upward water velocity needed to counteract the downward diffusive flux at the surface is thus

$$v = \frac{z_{0.1}}{t} = \frac{2.3 \sqrt{D}}{\sqrt{t}} \quad (\text{A5.13})$$

and the corresponding flow rate

$$q_g = \frac{2.3 p \sqrt{D}}{\sqrt{t}} \quad (\text{A5.14})$$

where p is the porosity.

The tracer change of the surface water decreases with an increasing rate of groundwater outflow, and becomes negligible at the outflow rate given by Eq.(A5.14). The corresponding condition for tracer change during infiltration is not obtained directly from Eq.(A5.3), since in this case the tracer front is moving downward. Due to lack of mixing in the soil water, the tracer concentration at the front is smaller than c_{in} . The diffusive flux at the front is less than that given by Eq.(A5.5), and the flux at the surface is still less. Tracer changes during overland flow on recharge areas, i.e., during Hortonian overland flow, are therefore disregarded.

In Fig.A5.5 the mean rates of groundwater flow per unit of saturated area in the present investigation, both estimated by

^{18}O , are compared with the outflow rate needed to prevent molecular exchange according to Eq.(A5.14) with $p = 0.5$ and $D = 1.68 \cdot 10^{-9} \text{ m}^2 \text{ s}^{-1}$. The figure shows that the outflow rates are usually large enough to prevent any considerable diffusive flux into the groundwater. But most snowmelt events fall close to the curve for q_g , and for some (small) snowmelt events the diffusive flux cannot be disregarded. The figure shows estimated mean rates of groundwater outflow in the discharge areas. Considering the variability in hydraulic conductivity and topographical conditions, there is a large variability in the rate of outflow within the discharge areas. If the outflow is concentrated in certain parts of a discharge area, the rate of outflow will be smaller in other parts, and the tracer change could be larger than that obtained from Eq.(A5.10) and the assumed values of Y'' .

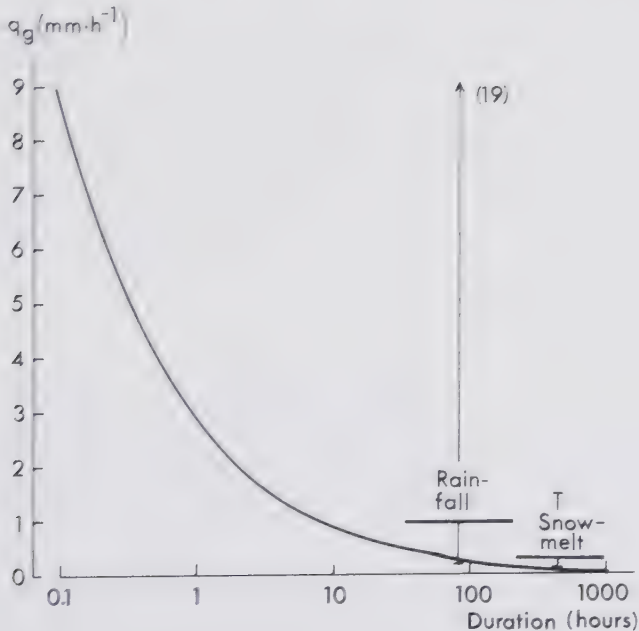


Fig.A5.5 Mean upward groundwater flow, q_g , needed to counteract the diffusive flux as a function of the duration of the flow situation. Median values and ranges of groundwater discharge per discharge area, as estimated by ^{18}O , and approximate durations of the two types of events are marked in the figure.

Conclusion and discussion on molecular exchange between overland flow and groundwater

In the above discussion it has been examined whether the isotope content of water flowing on the ground is likely to be changed by molecular exchange with the water in the soil on which it flows. Such a change would cause an overestimation of the groundwater fractions calculated by Eq.(3.3), since fresh rain- or meltwater reaching the stream isotopically would look partly like groundwater. It should be noted that the results of the theoretical considerations in this section are merely indicative, being based on incomplete field data.

It has been shown that the possibility of a modification of the isotope content of the overland flowing water cannot be excluded for all flow situations. The largest modification occurs at small values of q/Y , i.e., at small rates of overland flow per unit of saturated area with no groundwater outflow. For the present investigation, the effect can safely be disregarded at high-intensity rainfall events and probably also during medium-intensity rainfall and high-intensity snowmelt events. During low intensity snowmelt events, a modification affecting the chemical or isotope hydrograph separation is, however, not inconceivable. The modification may also occur at the early stages of higher-intensity events.

Since tracer changes are small at high-intensity events, the comparatively large groundwater fractions observed at such events, are accurate. However, the inverse relationship found between groundwater fraction and rate of water input (e.g. Fig.10.2) could be partly fictitious according to the present discussion.

The choice of q in Eq.(A5.10), and thus the estimated tracer change, depends on the assumed model for streamflow generation. In Table A5.1 q_1 represents the Horton model (with very little infiltration) while q_2 represents the present model. Fig.A5.3 shows that with the former values the tracer change is negligible. A possible modification with the latter values implies that the estimated fractions of discharge area are too small, since in reality more new rain- or meltwater reached the stream directly.

In judging the effect of diffusive tracer changes on hydrograph separation it has so far been assumed that the surface near groundwater in "no-flow areas" is isotopically identical with

the groundwater contributing to streamflow. But the topmost groundwater in such areas is more likely to consist of newly infiltrated water. Due to the very small or non-existing groundwater outflow, the upper layers dry by evapotranspiration after an event and refill by infiltration during the next. This is particularly true for "no-flow areas" within temporary discharge areas. If the groundwater is purely event-infiltrated water, there will be no tracer change of the overland-flowing water, of course. On the other hand, if the non-flowing groundwater to a considerable degree consists of pre-event rainwater differing isotopically from the rainfall of the event, the hydrograph separation might be distorted in either direction.

Finally, it should be noted that any molecular exchange that might occur only affects the results concerning the water-flow pattern, which is important for instance in the development of runoff models. From a chemical point of view, the discharged amount of rain- or meltwater exceeding the one estimated by ^{18}O is groundwater. This is so since the coefficient of diffusion for chemical constituents in water is about the same as the one for ^{18}O . If there is a considerable exchange of ^{18}O , there would also be correspondingly large exchange of the other constituents.

